GEOLOGICAL SURVEY OF CANADA
PAPER 89-10

THE GEOLOGY OF THE ANTIGONISH HIGHLANDS, NOVA SCOTIA

J. Brendan Murphy
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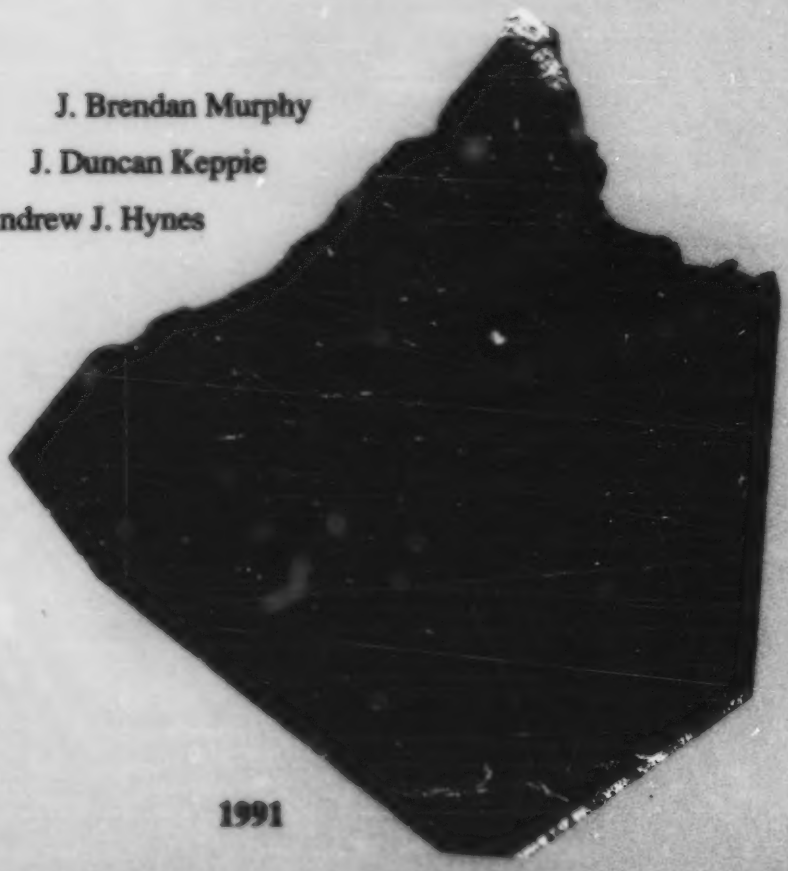

1991

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Contribution to Canada-Nova Scotia Mineral Development Agreement 1984-89, a subsidiary agreement under the Economic and Regional Development Agreement. Project funded by the Geological Survey of Canada.

Contribution à l'Entente auxiliaire Canada/Nouvelle-Écosse sur l'exploitation minérale 1984-89 faisant partie de l'Entente de développement économique et régional. Ce projet a été financé par la Commission géologique du Canada.

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92-01896/

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Available in Canada through

authorized bookstore agents and other bookstores

or by mail from

Canada Communication Group - Publishing
Ottawa, Canada K1A 0S9

and from

Geological Survey of Canada offices:

601 Booth Street
Ottawa, Canada K1A 0E8

3303-33rd Street N.W.,
Calgary, Alberta T2L 2A7

A deposit copy of this publication is also available for reference
in public libraries across Canada

Cat. No. M44-89/10E
ISBN 0-660-14214-7

Price subject to change without notice

Cover description

IHS (intensity, hue, saturation) transformed image of
LANDSAT MSS and airborne radar. Radar data from CCRS.
The image correlates well with the geology (see Figure 3).
The northeasterly trending Hollow Fault (top of picture),
the easterly trending Chedabucto Fault (bottom) and the
Eden Lake vent (?) complex stand out in particular.

Critical Reader
C. van Staal

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Original manuscript received: 1988-02-15
Final version approved for publication: 1988-02-19

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THE GEOLOGY OF THE ANTIGONISH HIGHLANDS, NOVA SCOTIA

Abstract

The Antigonish Highlands consist predominantly of late Precambrian rocks of the Georgeville Group. The stratigraphy of the Georgeville Group records the progressive development of a marine basin. Subaerial to shallow marine volcanic-dominated sequences are overlain by a thick sequence of turbidites with a volcanic arc source area. The geochemistry of the volcanic rocks indicates that there are at least three distinct contemporaneous volcanic rock types: (i) a calc-alkalic basaltic andesite; (ii) a continental within-plate tholeiitic-alkalic basalt; and (iii) volcanic arc-related felsic volcanic rocks. The geochemistry, stratigraphy, and geochronology suggest that the Georgeville Group is synorogenic, genetically associated with subduction, and was deposited in an intra-arc or peri-arc basin. Geochronological data constrain the depositional age to ca. 618-610 Ma. Deformation of the Georgeville Group by thrusts, easterly vergent isoclinal folds, and north-south and east-west upright folds is associated with westward telescoping and closure of the basin. Synchronous movement along major northeast-trending strike-slip faults may have accompanied this deformation and may have continued into the Cambrian. Intrusions by ca. 615-610 Ma appinitic gabbro and granite are spatially and temporally associated with movement along major faults.

Cambrian-Lower Ordovician rocks of the Antigonish Highlands were deposited in a pull-apart basin associated with a dextral transpressional stage of the late Precambrian Avalonian-Cadomian Orogeny. Two groups of rocks were deposited: the predominantly sedimentary Iron Brook Group and the predominantly volcanic McDonalds Brook Group. The rocks were deposited in a continental to shallow marine environment. The geochemistry of the volcanic rocks is consistent with a continental within-plate tectonic environment and is genetically related to the development of the pull-apart basin. Cambrian plutonic rocks (granite and gabbro) display geochemical characteristics consistent with a transpressional environment. The basin was deformed by thrusts and isoclinal folds probably in the Middle Ordovician. Deformation is possibly related to sinistral motion on northeast-trending faults.

Late Ordovician to Early Devonian rocks of the Arisaig Group consist of mafic to felsic volcanic rocks overlain by a thick sequence of fossiliferous siliciclastic rocks, and were deposited in a continental to shallow marine environment. The geochemistry of the volcanic rocks is consistent with a continental rift-related tectonic environment. These rocks were deformed by faults and thrusts, probably in the Middle Devonian. Upper Devonian-Carboniferous plutonic rocks are associated with the transpressional stages of the Acadian Orogeny and may be related to the docking of the Meguma Terrane.

Résumé

Les hautes-terres d'Antigonish sont principalement formées de roches de la fin du Précambrien appartenant au groupe de Georgeville dont la stratigraphie reflète la formation progressive d'un bassin marin. Une épaisse séquence de turbidites provenant d'une région d'arc volcanique recouvre des séquences dominées par des roches volcaniques mises en place dans un milieu subaérien à marin peu profond. La géochimie des roches montre qu'il existe au moins trois genres contemporains distincts de roches volcaniques : i) des andésites basaltiques calco-alcalines, ii) des basaltes tholéiitiques et alcalins à caractère continental formés au sein d'une même plaque et, iii) des roches volcaniques de nature felsique typiques d'un arc volcanique. La géochimie, la stratigraphie et la géochronologie du groupe de Georgeville semblent indiquer que le groupe serait d'origine syntectonique, aurait été mis en place dans un bassin intra-arc ou péri-arc et qu'il serait génétiquement associé à la subduction. Les données géochronologiques permettent d'établir que la sédimentation a eu lieu il y a 618 à 610 Ma. Le groupe de Georgeville a été déformé par des failles de poussée, des plis isoclinaux avec vergence vers l'est et des plis droits de direction nord-sud et est-ouest, tous associés à un chevauchement vers l'ouest et à la fermeture du bassin. Il est possible que des déplacements le long de décrochements majeurs de direction nord-est aient accompagné cette déformation et se soient poursuivis au cours du Cambrien. La répartition spatiale et temporelle d'intrusions de gabbro appinitique et de granite âgés d'environ 615 à 610 Ma est liée au mouvement le long de failles majeures.

La mise en place des roches des hautes-terres d'Antigonish du Cambrien et de l'Ordovicien inférieur a eu lieu dans des bassins de «pull-apart» (extension) associés à la transpression dextre de la phase tectonique avalonienne et cadomienne de la fin du Précambrien. Deux groupes de roches se sont alors formés : le groupe d'Iron Brook, principalement sédimentaire, et le groupe de McDonalds Brook, principalement volcanique. Les roches ont été mises en place dans un milieu continental à marin peu profond. La géochimie des roches volcaniques est compatible avec un milieu tectonique continental s'étant développé au sein d'une même plaque; l'origine de ces roches serait associée à l'agrandissement du basin de «pull-apart». Les caractéristiques géochimiques des roches plutoniques cambriennes (granite et gabbro) est compatible avec un milieu de transpression. Le bassin a été déformé par des chevauchements et des plis isoclinaux, probablement au cours de l'Ordovicien moyen; cette déformation pourrait être liée au déplacement senestre des failles de direction nord-est.

Les roches de l'Ordovicien supérieur au Dévonien inférieur formant le groupe d'Arisaig comprennent des roches volcaniques de nature mafique à felsique recouvertes d'une épaisse séquence de roches silico-clastiques fossilifères; elles ont été mises en place dans un milieu continental à marin peu profond. La géochimie des roches volcaniques est compatible avec un milieu tectonique lié à un rift continental. Des failles et des chevauchements sont venus déformer ces roches, probablement au Dévonien moyen. Les roches plutoniques de la fin du Dévonien et du Carbonifère sont associées aux transpressions de la phase tectonique acadienne et pourraient être liées à l'accolement du terrane de Meguma.

SUMMARY

This study has resulted in a complete re-evaluation of the geology of the Antigonish Highlands. The Antigonish Highlands lie in the Avalon Terrane of the Canadian Appalachians (Williams, 1979; Keppie, 1985b) and consist of late Precambrian to Early Devonian rocks. Late Precambrian rocks are assigned to the Georgeville Group and record the progressive development of a sedimentary basin. The oldest rocks of the Georgeville Group occur in the north and south of the highlands whereas the youngest rocks occur in the central highlands. The oldest rocks comprise subaerial to shallow marine bimodal volcanic rocks. These volcanic rocks consist of interlayered calc-alkalic basaltic andesites (with some tholeiitic mid-ocean ridge basalt (MORB) tendencies), continental tholeiitic-alkalic within-plate basalts and felsic volcanic rocks that were probably generated by anatexis of the crust associated with rising mafic magma. The geochemistry of the volcanic rocks is consistent with an intra-arc or peri-arc tectonic setting. These rocks are interlayered with volcanogenic and calcareous sediments and are overlain by a thick sequence of turbidites and minor mafic volcanics. If the Georgeville Group is penecontemporaneous with the arc-related late Precambrian Fourchu Group of southeastern Cape Breton Island, then the volcanism may be related to late Precambrian subduction associated with the Avalonian-Cadomian Orogeny. Geochronological data constrain the age of deposition of the Georgeville Group to ca. 618-610 Ma. Closure of the Georgeville Basin occurred in the late Precambrian, resulting in the deformation of the Georgeville Group by easterly vergent isoclinal folds and thrusts followed by upright north-south to northeast-southwest open folds. Greenschist facies metamorphism accompanied the isoclinal folding. Late Precambrian plutons consist of ca. 615-610 Ma appinitic gabbro, alaskite, and possibly granite. These plutons are posttectonic with respect to late Precambrian folding; however the intrusion of the appinites may be spatially and temporally associated with movement along major faults. The

SOMMAIRE

La présente étude a donné lieu à une réévaluation complète de la géologie des hautes-terres d'Antigonish. Ces hautes-terres font partie du terrane d'Avalon des Appalaches canadiennes (Williams, 1979; Keppie, 1985b) et sont formées de roches de la fin du Précambrien au Dévonien inférieur. Les roches de la fin du Précambrien sont assignées au groupe de Georgeville et témoignent de la formation progressive d'un bassin sédimentaire. On retrouve dans le nord et dans le sud des hautes-terres les roches les plus anciennes du groupe de Georgeville alors que les roches les plus récentes se trouvent dans la partie centrale de la région. Les roches les plus anciennes comprennent des roches volcaniques bimodales mises en place dans un milieu subaérien à marin peu profond. Ces roches sont interstratifiées avec des sédiments calcaires et d'origine volcanique et sont recouvertes d'une épaisse séquence de turbidites. Les roches volcaniques qui ont suivi sont composées d'andésites basaltiques calco-alcalines interstratifiées (montrant certaines caractéristiques propres aux basaltes tholéiitiques des dorsales médio-océaniques), de basaltes tholéiitiques et alcalins à caractère continental s'étant formés au sein d'une même plaque et de roches volcaniques felsiques résultant probablement d'une anatexie de la croûte associée à une montée de magma mafique. La géochimie des roches volcaniques est compatible avec un milieu tectonique intra- ou péri-arc. Si le groupe de Georgeville et le groupe de Fourchu, datant de la fin du Précambrien et situé dans la partie sud-est de l'île du Cap-Breton, sont pénécontemporains, le volcanisme pourrait alors être lié à la subduction qui s'est produite à la fin du Précambrien, au cours de la phase tectonique avalonienne-cadomienne. Les données géochronologiques permettent d'établir que la sédimentation a eu lieu il y a 618 à 610 Ma. La fermeture du bassin de Georgeville s'est produite à la fin du Précambrien, menant à une déformation du groupe de Georgeville par des plis isoclinaux avec vergence vers l'est, des chevauchements et, plus tard, par des plis droits ouverts de direction nord-sud à nord-est-sud-ouest. La formation des plis isoclinaux a été accompagnée de métamorphisme à faciès des schistes verts. Les plutons de la fin du Précambrien se composent de gabbro appinitique, d'alaskite et possiblement de

limited time gap between deposition, deformation, and intrusion of the Georgeville Group indicates that the Georgeville Group was probably deposited synorogenically in a strike-slip basin within an ensialic volcanic arc regime.

Cambrian-Early Ordovician rocks unconformably overlie the Georgeville Group. The contact is an unconformity in some localities; in others it is a low-angle fault. Cambrian-Early Ordovician rocks occur only in the northernmost Antigonish Highlands and are divided into two groups that are related by lateral facies variations: the predominantly sedimentary Iron Brook Group, and the predominantly volcanic McDonalds Brook Group. The present contacts between these two groups are thrusts. The Iron Brook Group consists of red fluviatile conglomerates and red slates overlain by fossiliferous (middle Lower Cambrian) pink limestones, red slates, and calcareous tuffs, followed by ironstones and quartzites which may be Early Ordovician in age. An unconformity or diastem may occur within the Iron Brook Group. The McDonalds Brook Group consists of red fluviatile conglomerates and shales at the base overlain by a thick sequence of bimodal volcanic rocks with minor interlayered red clastic sediments and pink limestones. The stratigraphy suggests a transition from a subaerial to shallow marine environment of deposition. The mafic rocks are alkalic, within plate, extensional, and are interpreted to be related to local rifting. Felsic volcanic rocks are attributed to crustal anatexis associated with ascending mafic magma. The limited occurrence of Cambro-Ordovician sequences within Nova Scotia and the fault-bounded nature of the Antigonish sequence, together with the geochemistry of the volcanic rocks, indicates a local pull-apart basin tectonic environment probably associated with dextral transpression at the latest stages of the Cadomian Orogeny. The presence of a cosmopolitan Acado-Baltic fauna in the pink limestones indicates a connection to open seas. Cambrian-Early Ordovician rocks were deformed, probably in the Middle Ordovician, by isoclinal folds, thrusts, and upright north-south folds associated with sinistral transpression. The deformation was accompanied by greenschist facies metamorphism, which is most intense in the centre of the basin and dies out toward the margins. This is one of the few documented examples of Ordovician deformation in the Avalon Terrane. The underlying Precambrian rocks are virtually unaffected by this deformation indicating that Precambrian and Cambrian-Early Ordovician sequences are separated by a décollement surface that lies close to, or at the contact between Precambrian and Cambrian rocks.

Late Ordovician-Early Devonian rocks are assigned to the Arisaig Group. These rocks unconformably overlie Cambrian to Early Ordovician rocks. They predominantly occur at the extremities of the Antigonish Highlands. They consist of fluviatile red clastic sediments and interlayered bimodal volcanic rocks overlain by a thick sequence of fossiliferous, shallow marine siliciclastic rocks and minor rhyolites with red clastic sediments toward the top of the sequence. The stratigraphy records a marine incursion at the base of the Silurian, which occurred possibly in response to a worldwide eustatic rise in sea level.

granite âgés d'environ 615 à 610 Ma. Il s'agit de plutons post-tectoniques par rapport au plissement de la fin du Précambrien; cependant, la répartition spatiale et temporelle des intrusions de gabbro appinitque est peut-être liée au mouvement le long de failles majeures. Le court intervalle de temps qui s'est écoulé entre la sédimentation, la déformation et l'intrusion du Groupe de Georgeville indique qu'il s'est probablement agi d'une sédimentation syntectonique survenue dans un bassin créé par un décrochement au sein d'un régime d'arc volcanique ensialique.

Les roches du Cambrien et de l'Ordovicien inférieur reposent en discordance sur le groupe de Georgeville. À certains endroits, le contact se présente sous forme de discordance; ailleurs, il est marqué par une faille suhorizontale. Les roches du Cambrien et de l'Ordovicien inférieur ne se trouvent que dans la partie la plus septentrionale des hautes-terres d'Antigonish et forment deux groupes liés par des variations latérales de faciès, soit le groupe d'Iron Brook, principalement sédimentaire, et le groupe de McDonalds Brook, principalement volcanique. Les contacts actuels entre ces deux groupes sont des failles chevauchantes. Le groupe d'Iron Brook se compose de conglomérats fluviatiles rouges et de schistes ardoisiers rouges, recouverts de calcaires roses fossilifères (milieu du Cambrien inférieur), de schistes ardoisiers rouges et de tufs calcaires sur lesquels reposent des roches sédimentaires ferrugineuses et des quartzites pouvant dater de l'Ordovicien inférieur. Il semble qu'une discordance ou lacune stratigraphique se manifeste au sein du groupe d'Iron Brook. À la base du groupe de McDonalds Brook se trouvent des shales et des conglomérats fluviatiles rouges; ces roches sont recouvertes par une épaisse séquence de roches volcaniques bimodales interstratifiées avec de petites quantités de sédiments clastiques rouges et de calcaires roses. La stratigraphie laisse supposer une transition d'un milieu sédimentaire subaérien à un milieu marin peu profond. Les roches mafiques sont alcalines et sont typiques d'un milieu de distension se produisant au sein d'une même plaque; on croit qu'elles sont liées à la formation locale de rifts. Les roches volcaniques de nature felsique pourraient résulter d'une anatexie de la croûte associée à une montée de magma mafique. La présence restreinte de séquences cambro-ordoviciennes en Nouvelle-Écosse, le fait que la séquence d'Antigonish est délimitée par des failles et la géochimie des roches volcaniques, montrent que le milieu tectonique local prend la forme d'un bassin de «pull-apart» probablement associé à une transpression dextre qui s'est produite au cours des dernières étapes de la phase tectonique cadomienne. La présence dans les calcaires roses d'une faune cosmopolite de nature acadienne et baltique dénote la présence d'un lien avec la haute mer. Probablement au cours de l'Ordovicien moyen, les roches du Cambrien et de l'Ordovicien inférieur ont été déformées par des plis isoclinaux, des chevauchements et des plis droits de direction nord-sud associés à une transpression senestre. Cette déformation était accompagnée de métamorphisme à faciès des schistes verts, faible en bordure du bassin et s'intensifiant vers le centre. Il s'agit d'un des rares exemples documentés de déformation ordovicienne dans le terrane d'Avalon. Les roches précambriennes sous-jacentes ne sont pratiquement pas touchées par cette déformation, montrant que les séquences précambriennes et celles du Cambrien et de l'Ordovicien inférieur sont séparées en raison d'un décollement au contact entre les roches précambriennes et cambriennes, ou tout près.

The geochemistry provides evidence that the marine incursion was accompanied by local rifting. Late Ordovician-Early Silurian volcanic rocks consist of continental, alkalic-tholeiitic, within-plate extensional mafic rocks and rhyodacitic felsic rocks that were probably generated by crustal anatexis associated with the ascending mafic magma. Rhyolites may be related to the rhyodacites by differentiation. This magmatism is synchronous with the early Silurian docking of the Avalon terrane to North America. Deformation of the Arisaig Group is heterogeneous, with the rocks varying from relatively undeformed in many areas to intensely deformed in association with shear zones, isoclinal folds, and thrusts adjacent to the Chedabucto Fault. The age of this deformation is uncertain, but at least some of it is probably Middle Devonian and may be related to movement along the Chedabucto Fault. A Carboniferous age for some of this deformation cannot be ruled out. Faults of unknown age offset Devono-Carboniferous rocks.

The tectonic environment for the Georgeville Group indicates that it has the potential for Cu-Au-Mo-Hg, Zn-Cu-Pb, and Kuroko-type deposits. Several interesting pyritic zones were found, some with considerable width and strike-length in the vicinity of mineralized stockwork veining in volcanic breccia. This suggests that the Keppoch Formation may have the potential for epithermal deposits. The tectonic environment for Cambrian-Early Ordovician and Late Ordovician rocks indicates a potential for Sn-W-Mo-F-U-Ag-Nb deposits. Promising pyritic zones are associated with the intrusion of felsic "quartz-eye" porphyry of unknown age, occurring adjacent to the Browns Mountain Fault. A vent complex west of Eden Lake, associated with early Silurian volcanism may also have considerable potential.

Les roches de l'Ordovicien supérieur et du Dévonien inférieur font partie du groupe d'Arisaig; elles reposent en discordance sur les roches sous-jacentes, qui datent du Cambrien et de l'Ordovicien inférieur, et se trouvent principalement aux extrémités des hautes-terres d'Antigonish. Ces roches se composent de sédiments clastiques rouges d'origine fluviatile et de couches interstratifiées de roches volcaniques bimodales, le tout sous-jacent à une épaisse séquence de rhyolites peu abondantes et de roches silico-clastiques fossilifères mises en place dans un milieu marin peu profond; on retourve des sédiments clastiques rouges dans la partie supérieure de la séquence. La stratigraphie témoigne d'une transgression marine au début du Silurien, possiblement amenée par une hausse eustatique mondiale du niveau de la mer. La géochimie montre que cette trangression marine était accompagnée à certains endroits par la formation de rifts. Les roches de l'Ordovicien supérieur et du Silurien inférieur se composent de roches mafiques de nature alcaline et tholéiitique, à caractère continental, s'étant formées au sein d'une même plaque et ayant été mises en place dans un milieu de distension, et de roches felsiques rhyodacitiques probablement formées par anatexie de la croûte lors de la montée de magma mafique. Les rhyolites pourraient être associées aux rhyodacites par différenciation. L'épisode de magmatisme a coïncidé avec l'accolement du terrane d'Avalon à l'Amérique du Nord au début du Silurien. Le groupe d'Arisaig est très déformé aux alentours de zones cisaillées, de zones de chevauchement et de plis isoclinaux adjacents à la faille Chedabucto, alors qu'il est relativement non déformé à de nombreux endroits. L'âge de cette déformation est incertain; on pense qu'elle s'est probablement produite, du moins en partie, au Dévonien moyen et qu'elle pourrait être liée à un déplacement le long de la faille Chedabucto. Elle pourrait aussi, en partie, dater du Carbonifère. Des failles d'âge inconnu décalent des roches datant du Dévonien et du Carbonifère.

Le milieu tectonique du groupe de Georgeville indique que ce dernier pourrait bien renfermer des gisements de Cu-Au-Mo-Hg et de Zn-Cu-Pb, ainsi que des gisements de type Kuroko. Plusieurs zones pyriteuses intéressantes ont été relevées, dont certaines, aux alentours de réseaux de fissures minéralisées dans la brèche volcanique, sont de largeur et de direction et longueur considérables. Un tel phénomène laisse supposer que la formation de Keppoch pourraient renfermer des gisements épithermaux. Le milieu tectonique des roches du Cambrien et de l'Ordovicien inférieur et de l'Ordovicien supérieur dénote la présence possible de gisements de Sn-W-Mo-F-U-Ag-Nb. Des zones pyriteuses prometteuses sont associées à l'intrusion de porphyre felsique «à oeil de chat» d'âge inconnu, laquelle est adjacente à la faille Browns Mountain.

INTRODUCTION

The Antigonish Highlands are located in northeastern mainland Nova Scotia between the latitudes of 45°22′ and 45°52′ north and the longitudes of 61°55′ and 62°40′ west (Fig. 1). The area is about 1500 km^2 in size and includes western Antigonish County, eastern Pictou County, and northernmost Guysborough County.

The Antigonish Highlands are bound to the north and west by the Hollow Fault, to the south by the Chedabucto Fault and to the east by the Devono-Carboniferous Antigonish Basin. The highlands consist predominantly of late Precambrian rocks (Georgeville Group) but also contain sequences of Cambrian-Early Ordovician age (McDonalds Brook and Iron Brook groups) and latest Ordovician-Early Devonian age (Arisaig Group). Plutonic rocks, predominantly consisting of gabbro and granite, occur throughout the highlands. These intrusive rocks range in age from late Precambrian to Devono-Carboniferous. The late Precambrian-Early Devonian rocks are generally more resistant to erosion than the surrounding Devono-Carboniferous sedimentary rocks in the Antigonish and Stellarton basins. The Antigonish Highlands lie in the Avalon Zone (Williams, 1978) or Avalon "Composite" (Keppie, 1982, 1985b) or "Suspect" (Williams and Hatcher, 1982) Terrane of the Canadian Appalachians. The diversity of Late Precambrian successions in the Avalon Terrane contrasts with the Cambro-Ordovician stratigraphy which may be traced across the entire terrane (Keppie, 1985). Other areas that form part of the Avalon Composite Terrane in Nova Scotia include the Cobequid Highlands to the west, and most, if not all of Cape Breton Island to the east, (compare Barr and Raeside, 1989, 1990; Keppie, 1985b, 1990). Precambrian rocks of the Antigonish Highlands may be correlated with those of the Cobequid Highlands (Murphy et al., in press), but, in detail are atypical of Avalonian rocks. On the other hand, the Cambrian-Early Devonian rocks of this area are similar to those elsewhere in the Avalon Composite Terrane. The collision of these Precambrian terranes is recorded by the latest Precambrian Avalonian-Cadomian deformation (Keppie, 1982).

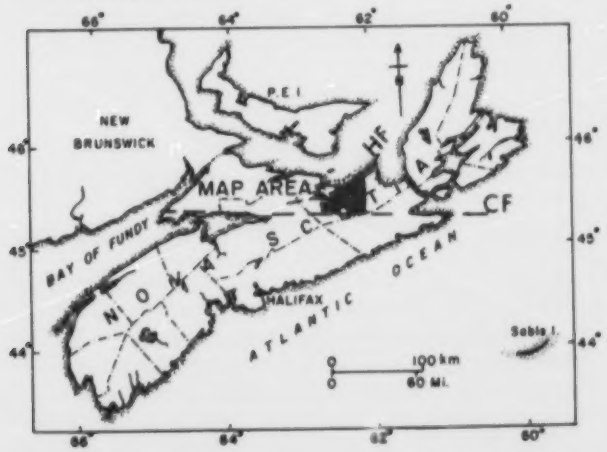


Figure 1. Location map of the Antigonish Highlands in Nova Scotia. A map of the darkened area is presented in Figure 3. HF-Hollow Fault, CF-Chedabucto Fault.

Previous work

The earliest geological maps of the Antigonish Highlands were prepared by Fletcher (1887, 1892) at a scale of 1 inch to 1 mile. On these maps, the Silurian-Carboniferous rocks were identified and the plutons were outlined. However, large areas of the highlands were not subdivided and were presumed to be of Cambro-Ordovician age. The northern Antigonish Highlands were mapped by Williams (1914), with special emphasis placed upon the Silurian rocks. Nevertheless, Williams (1914) did distinguish pre-Silurian areas consisting of greywacke, slate, and ironstone (James River Formation) from those consisting of red and grey sandstone and slate (Baxter Brook Formation) and rhyolite, basalt, and conglomerate (Malignant Cove Formation). Williams (1914) included the James River and Baxter Brook formations in the Browns Mountain Group, which was deformed into upright folds and metamorphosed prior to the deposition of the Malignant Cove Formation. All of these rocks were presumed to be of Ordovician age based upon the occurrence of *Obolus (Lingulobus) spissa* and *Lingulella*(?) in the ironstone and associated sandstone.

Maehl (1961) published a detailed account of the Silurian stratigraphy of Pictou County. He assigned these rocks to the Arisaig Group and compared them with the type section at Arisaig. Maehl (1961) concluded that the Silurian rocks of northern Pictou County compared favourably with the type area for the Arisaig Group, whereas the rocks of southern Pictou County were significantly different from, but related to, the rocks of northern Pictou County by lateral facies variations.

More recently Boucot et al. (1974) provided a detailed account of the stratigraphy, paleontology, and distribution of the Siluro-Devonian rocks around Arisaig and the Devono-Carboniferous rocks around the shores of Cape George. These authors found a similar stratigraphy to that proposed by Maehl (1961). At about the same time, Benson (1967, 1970, 1974) systematically mapped the Antigonish Highlands and the surrounding areas. Benson concluded that most of the highlands consisted of the Browns Mountain Group. He divided the group into four formations, which are, from oldest to youngest: (i) the Keppoch Formation, consisting mainly of felsic volcanic rocks with quartzite, siltstone, and phyllite; (ii) the Baxter Brook Formation, containing tuffaceous siltstone, shale, and argillite, which he interpreted as the facies equivalent of part of (iii) the Brierly Brook Formation, made up predominantly of volcanic rocks with minor greywacke; overlain by (iv) the Little Hollow Formation, consisting mainly of red and grey siltstone and quartzite with minor ironstone and ferruginous wacke. Benson (1974) also concluded that: (i) the Browns Mountain Group was Cambro-Ordovician in age and was deformed during the Taconian Orogeny; (ii) it was then uplifted and eroded before being unconformably overlain first by the Malignant Cove Formation followed by the Siluro-Devonian rocks; and (iii) all of these rocks were deformed during the Acadian Orogeny, were uplifted and eroded and then were unconformably overlain by Devono-Carboniferous rocks.

During 1976 and 1977, detailed work in the Arisaig and Georgeville areas and a reconnaissance of the entire Antigonish Highlands were carried out, and several significant discoveries were made (Keppie, 1978). Firstly, early recumbent folds were discovered in the Browns Mountain Group. These folds predated the intrusion of a posttectonic alaskite stock yielding a Rb/Sr whole rock isochron age of 535 ± 13 Ma (R.F. Cormier, pers. comm., 1981) and a K/Ar age on muscovite of 604 ± 14 Ma (R.K. Wanless, pers. comm., 1978). Secondly, a rich fauna was discovered in the limestones associated with the ironstones. This fauna was subsequently determined to be Early Cambrian in age (Landing et al., 1980). The implications of these observations are as follows:

1) At least two groups of rocks exist in what was formerly called the Browns Mountain Group. The older group is called the Georgeville Group and is Precambrian in age, whereas the younger fossiliferous group is clearly Phanerozoic, probably early Paleozoic, in age (Keppie, 1978). The Cambro-Ordovician rocks were assigned to the Iron Brook and McDonalds Brook groups (Murphy et al., 1980).

2) The recognition of early recumbent folds in the Georgeville Group, where none were previously identified, means that the stratigraphy and map patterns proposed by previous authors are suspect. Consequently, the stratigraphy and map patterns are in need of revision to be consistent with the polyphase structure. The early recumbent folds and subsequent east-west upright folds were correlated with the late Precambrian Cadomian Orogeny (Keppie, 1978).

3) These problems, together with the almost total lack of exposure in the Browns Mountain area, led Keppie (1978) to recommend abandoning the term Browns Mountain Group.

Aeromagnetic survey

During 1977 and 1978, the Antigonish Highlands and the surrounding areas were flown at 300 m above the ground at 300 m line spacings with a gradiometer mounted on a Queenair aircraft, by the Geological Survey of Canada as part of the DREE 1974-9 program (Geological Survey of Canada, 1982). This technique uses two high sensitivity magnetometers mounted ~3 m vertically apart and produces two types of magnetic maps: total field, and vertical gradient. The vertical gradient map is an extremely useful mapping tool (Hood et al., 1979) because (i) the magnetic gradient is produced mainly by the near surface rocks, and (ii) at this latitude, the zero contour approximately coincides with the geological boundary between magnetic and less magnetic units when the boundary is steeply dipping. These maps were published as magnetic contours on a topographic base and also with the magnetic data coloured and superimposed upon a topographic base with geological boundaries marked (Geological Survey of Canada, 1982). These maps, especially the vertical gradient maps, proved to be invaluable in locating geological boundaries inland where exposure is limited.

Magnetic units in the area causing positive magnetic anomalies include: (i) mafic volcanic rocks in the Precambrian Chisholm Brook and Clydesdale formations, in the Cambrian McDonalds Brook Group and in the Upper Ordovician-Lower Silurian Bears Brook Formation; (ii) felsic volcanic rocks in the McDonalds Brook Group; (iii) mafic intrusive igneous bodies; and (iv) some specific slate horizons in the Precambrian Maple Ridge Formation. One feature of particular interest is the concentric magnetic anomaly pattern in Silurian rocks to the west of Eden Lake. This may represent a vent complex. Magnetite appears to be the main magnetic mineral in the volcanic and plutonic igneous lithologies, and magnetite is also present in the magnetic slates of the Maple Ridge Formation. In most cases, the geological boundaries are steeply dipping. Judicious use of outcrop lithologies, strike and dip data, the vertical gradient maps and air photo interpretation facilitated the drawing of geological boundaries where contacts were not observed in the field.

Purpose

The purpose of this study is to re-examine the geology of the Antigonish Highlands. This involves:

1) mapping the distribution of the pre-Upper Devonian rocks of the Antigonish Highlands

2) determining the stratigraphic sequence

3) analyzing the structural and metamorphic histories

4) describing the geochemistry and evaluating the petrogenesis of the igneous rocks

5) determining the tectonic history of the highlands and evaluating its role in the evolution of the Appalachian Orogenic Belt

6) evaluating the economic potential of the highlands.

Organization of the report

Three major stratigraphic sequences may be recognized in the Antigonish Highlands: the late Precambrian Georgeville Group, the Cambrian-Lower Ordovician McDonalds Brook and Iron Brook groups and the uppermost Ordovician-Lower Devonian Arisaig Group. Each of these sequences is separated by an exposed angular unconformity, and there are fundamental differences in their geological histories and tectonic environments. Each of these sequences is described separately below in terms of its stratigraphy, depositional environment, magmatic history, and structural geology, and a model is proposed for the tectonic setting of each.

Acknowledgments

The authors acknowledge field support in the northern Antigonish Highlands by the Canada Department of Regional Economic Expansion and the Nova Scotia Department of Mines and Energy, and in the southern Antigonish Highlands by the Canada-Nova Scotia Co-operative Mineral Program 1981-1984 and the Canada-Nova Scotia Mineral Development Agreement 1984-1989, carried out by the Geological Survey of Canada.

Some of the analytical work reported here was funded by operating grants to J.B.M. and A.J.H. from the Natural Sciences and Engineering Research Council (NSERC) Canada. Special thanks to Fred Chandler for organizational support and to Alan Anderson, Dave Benson, Bob Boehner, Fred Chandler, Randy Cormier, Ken Currie, Dave Dallmeyer, Ken Galli, Wes Gibbons, Peter Giles, Jana Horak, Cynthia LaRocque, Damian Nance, Ken Northcote, Georgia Pe-Piper, David Piper, Bill Poole, Gao Ruixang, Al Sangster, Sid Taylor, Hau Chong Teng, Arden Thompson, Joe Whalen, and Gary Yeo for discussions with the authors in the field. Ken Northcote and Al Sangster are particularly thanked for sharing their insights on the mineralization in the area, Dave Dallmeyer and Tom Krogh for their excellent geochronological data, and to Jarda Dostal for his geochemical/petrological expertise. We were fortunate to have excellent field assistants: Roger Holstein (1978), Brian Fisher (1979), Mike Gurney (1983), Janet Campbell, Donnie Cameron, Mike Gillis and Wayne McNeil (1984), Kim Bellefontaine, Pat Campbell, Dave Hatchette, Eric Myketyn and Sandy Sears (1985). We would like to thank Marie Gillis, R. Jones, and S. Saunders for typing many (seemingly endless) versions of this manuscript. We are very grateful to Charlie Hickey, Maria MacIsaac, Sid Taylor and the Drafting Section of the Nova Scotia Department of Mines and Energy and the Geological Survey of Canada for their expertise in preparing the figures, and to Charlie for the overtime he put in to make sure this paper was submitted on schedule and to Helen for allowing Charlie to stick to his task. The manuscript was significantly improved by the insightful and constructive comments of Cees van Staal. Various parts of the manuscript were also improved by comments and suggestions from Sandra Barr, Kevin Cameron, Fred Chandler, Randy Cormier, Ken Currie, Richard D'Lemos, Jarda Dostal, Don Francis, Les Fyffe, Jeff Harris, Mike Hibbard, Cynthia LaRocque, Bob Martin, Alexander McBirney, Damian Nance, Georgia Pe-Piper, David Piper, Ron Pickerill, Bill Poole, Rob Raeside, Nick Rast, Andy Rencz, Klaus Schultz, Bob Stern, Rob Strachan, Sid Taylor, Hau Chong Teng, Anthony Williams-Jones, and Gary Yeo.

LATE PRECAMBRIAN ROCKS

Stratigraphy

Introduction

Late Precambrian rocks in the Antigonish Highlands are assigned to the Georgeville Group (Keppie, 1978). The Georgeville Group constitutes most of the Antigonish Highlands and outcrops in four fault-bounded blocks. From north to south these area: (i) the Georgeville block; (ii) the Maple Ridge block; (iii) the Clydesdale block; and (iv) the Keppoch block. A map showing the place names and the locations of the type sections is shown in Figure 2. A simplified map of the geology of the Antigonish Highlands is shown in Figure 3.

The stratigraphic units in the Maple Ridge and Clydesdale blocks are similar. However, the differences between these blocks and the Georgeville block are significantly enough to define separate formations in the Georgeville block. The type section of three formations of the Georgeville Group are located on the shoreline of the Northumberland Strait in the Georgeville area between Malignant Cove and Livingstone Cove. Inland, individual river sections were chosen as type sections for other formations (Fig. 2). The formations recognized in each block and their correlatives are indicated in Figure 3. Neither the stratigraphic base nor the top of the Georgeville Group is exposed in the Antigonish Highlands.

Age

Except for the recovery of an unidentifiable primitive sphaeromorph from the black mudstones of the Livingstone Cove Formation (J. Utting, pers. comm., 1977), no fossils have been found in the Georgeville Group. This is despite the numerous shale and mudstone samples analyzed for palynomorphs and several limestone samples dissolved for conodonts. Thus, the Precambrian age of the Georgeville Group must be established by other methods. A lower age limit is given by the 613 ± 5 Ma U-Pb age on detrital zircons from the Livingstone Cove Formation of the Georgeville Group (Keppie and Krogh, 1990). An upper age limit is established by the following relationships:

1) In the Georgeville block, the Chisholm Brook and Morar Brook formations were posttectonically intruded by the late Precambrain Greendale Complex (Fig. 3). Hornblendes from the Greendale Complex yield $^{40}Ar/^{39}Ar$ plateau ages of 620.3 ± 4.8 Ma and 611.2 ± 3.8 Ma (Keppie et al., 1990). In the Keppoch block, the late Precambrian Black Brook and Eden Lake complexes intrude the Keppoch Formation. Hornblendes from the Black Brook Complex yield a plateau age of 602.5 ± 3.2 Ma (Keppie et al., 1990). A K-Ar determination on biotite in the Eden Lake Complex yielded an age of 582 ± 32 Ma (Wanless et al., 1967). In the absence of other isotopic data, and assuming that there is little or no excess argon, it is suggested that these ages represent cooling ages and provide a minimum age on the time of intrusion. An alaskitic granite in the Georgeville block (Fig. 4) yields 604 ± 14 Ma (K-Ar on muscovite, R.K. Wanless, pers. comm., 1979), and 535 ± 13 Ma (Rb/Sr whole rock isochron, R.F. Cormier, pers. comm., 1981).

2) The Georgeville Group is unconformably overlain by the Eocambrian-Early Cambrian rocks that occur at the base of the McDonalds Brook and Iron Brook groups. The unconformity is exposed at Doctors Brook (Fig. 2 and 3). Although no angular discordance is visible at Doctors Brook, rhyolitic tuff at the base of the north-facing Iron Brook Group overlies green slates of the Maple Ridge Formation. The stratigraphic position of the Maple Ridge Formation within the Georgeville Group suggests that this is a major regional unconformity. Near Malignant Cove, conglomerates of the Lower Cambrian Malignant Cove Formation are inferred to rest unconformably on the Chisholm Brook Formation of the Georgeville Group.

In summary, geochronological data and field relationships constrain the depositional age of the Georgeville Group from ca. 618-610 Ma.

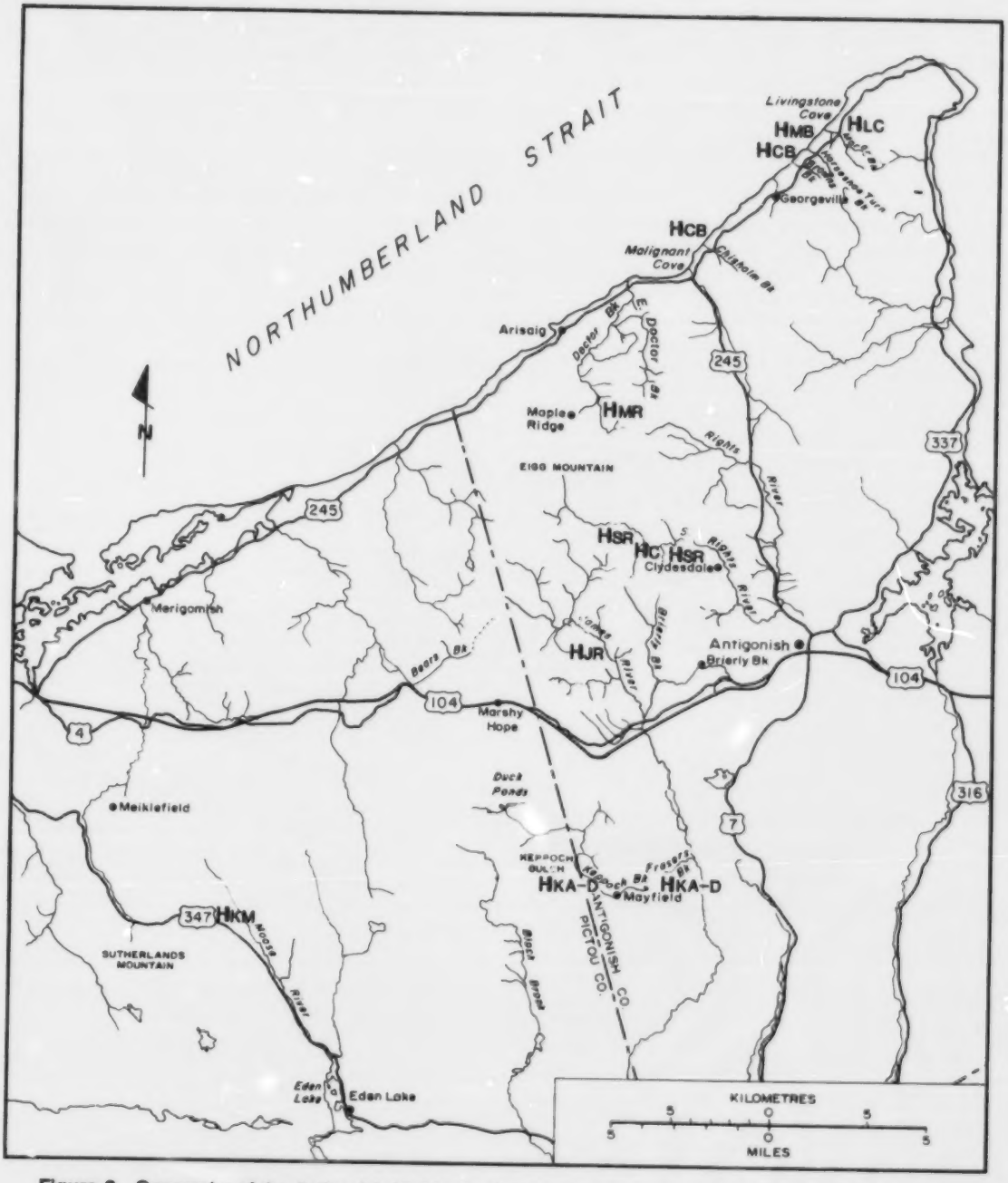
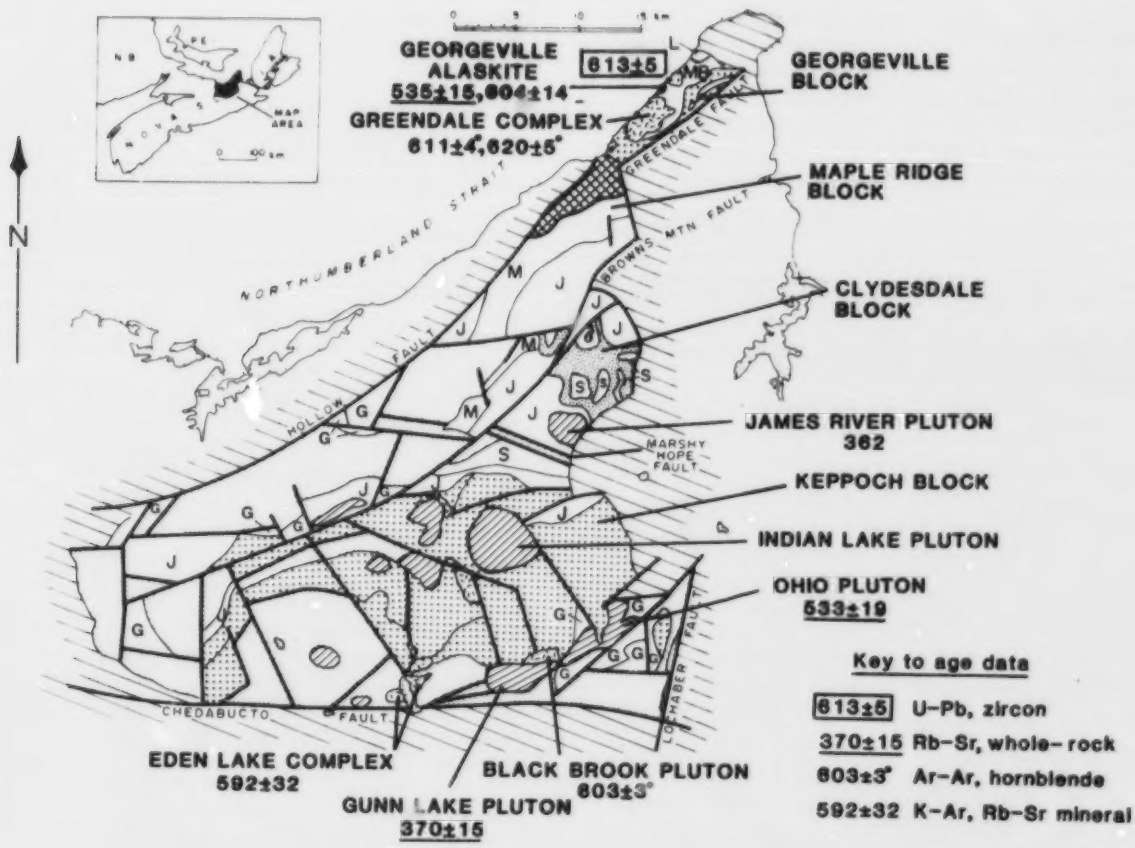


Figure 2. Geography of the Antigonish Highlands showing place names and locations of type sections. 1 = Chisholm Brook Formation; 1a = Keppoch F, Fraser Brook Member; 1b = Keppoch Formation, Moose River Member; 2 = Morar Brook Formation; 2a = Maple Ridge Formation; 3 = Livingstone Cove Formation; 3a = James River Formation; 4 = Clydesdale Formation; 5 = South Rights Formation.



Legend and Correlation Chart

	GEORGEVILLE BLOCK	MAPLE RIDGE & CLYDESDALE BLOCK	KEPPOCH BLOCK
PALEOZOIC	Undivided Paleozoic rocks Late Ordovician-Early Devonian rocks Cambro-Ordovician rocks		
PRECAMBRIAN	G GEORGEVILLE GROUP (Undivided)		
		S South Rights Fm	S South Rights Fm
		Clydesdale Fm.	
	L Livingstone Cove Fm.	J James River Fm.	J James River Fm
	MB Morar Brook Fm.	M Maple Ridge Fm.	
	>> Chisholm Brook Fm.		AA Moose R. Mbr. } Keppoch Fm. VV Frasers Bk. Mbr. }
PLUTONIC ROCKS	Granitoid rocks Gabbroic rocks		

Figure 3. Summary of the geology of the Antigonish Highlands. For details see Map 1749A, in pocket. The Georgeville block is bound by the Hollow and Greendale Faults. The Maple Ridge block is bound by the Greendale and Browns Mountain Faults. The Keppoch block is bound by the Browns Mountain and Chedabucto Faults.

Keppoch Formation

Field relationships. The Keppoch Formation occurs only within the Keppoch block of the southern Antigonish Highlands. It may be subdivided into two members based on the absence (Frasers Brook Member) or presence (Moose River Member) of finely laminated mudstones interlayered with the volcanic rocks. The type section of the Frasers Brook Member is exposed on Frasers Brook, with other good exposures on logging roads to the south of Frasers Brook. It is most easily accessible on logging roads in the Duck Ponds area (Fig. 2).

The Frasers Brook Member forms the core of the southern Antigonish Highlands and is flanked by successively younger rocks (i.e. the Moose River Member and the James River Formation). The contacts between the Frasers Brook Member and these units are exposed, allowing the age relationships to be deduced. The Frasers Brook Member is subdivided into sequences dominated by rhyolite, basaltic andesite and basalt, though it is important to note that these distinctions are made on a percentage basis and that interlayering of these lithologies occurs on all scales. The base of the Member is not exposed. However, the upper contact of the Frasers Brook Member with the Moose River member is defined by the first appearance of finely laminated mudstone interlayered with the volcanic rocks. Typically, the finely laminated mudstone unit in the vicinity of the contact is 1 to 5 m in thickness. The Frasers Brook Member is extremely variable in thickness but consists of at least 2200 m of interlayered felsic to mafic volcanic rocks in the Frasers Brook area.

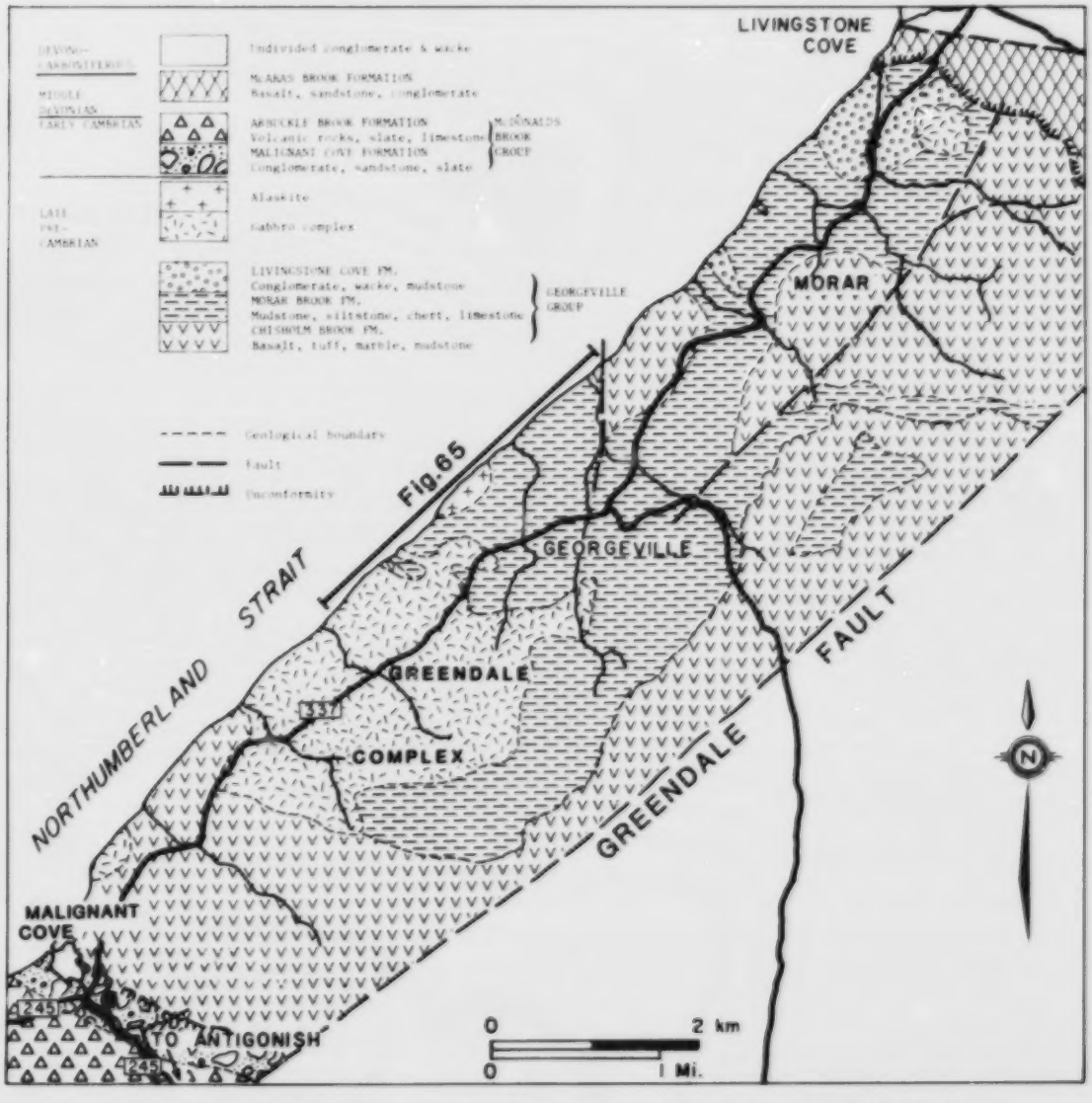


Figure 4. Summary of the geology of the Georgeville block, northern Antigonish Highlands.

The Moose River Member has its type section on Moose River, and is well exposed on logging roads to the east of Moose River and in the Meiklefield area. The contact of this member with the overlying James River Formation is placed where volcanic rocks are volumetrically minor. The contact is locally conformable in the western part of the Keppoch block, where the Moose River Member is 800 to 1000 m thick. However, in the eastern part of the Keppoch block, the James River Formation oversteps the Moose River Member and directly overlies the Frasers Brook Member.

Field description. The Keppoch Formation consists of felsic volcanic rocks (ignimbrites, tuffs and flows), basaltic andesites, basalts, volcanogenic sedimentary rocks (such as volcanic breccias and lahars) and mudstones. Felsic volcanic rocks are massive in texture. In the Frasers Brook Member they are pink, red or pale grey in colour, whereas in the Moose River Member they are generally pale green. Ignimbrites probably compose about 50 percent of the felsic volcanic rocks. They are typically aphanitic or contain up to 10 percent plagioclase and/or quartz phenocrysts, and display pale to dark red banding which is generally subparallel to bedding. Tuffs compose about 40 percent of the felsic volcanic rocks and contain felsic lithic fragments up to 3 cm in length. They are commonly fine grained and their tuffaceous aspect is visible only in thin section. Flows are aphyric or contain phenocrysts of plagioclase or quartz. They may be distinguished from the ignimbrites by their lack of banding, but can only be reliably distinguished from the fine grained tuffs in thin section.

Basaltic andesites are grey, green-grey or pale green in colour and contain plagioclase phenocrysts up to 3 cm in length. The matrix is aphanitic. Basalts may be distinguished from basaltic andesites by the paucity of plagioclase phenocrysts. Basalts are massive, green to dark green and aphanitic. Rarely, they contain phenocrysts of augite. They may also contain vesicles or quartz-chlorite filled amygdules. They are commonly spatially associated with mafic dykes and are best exposed in the Duck Ponds and Black Brook areas.

Lahars occur only in the Frasers Brook Member. They are composed of a distinctive purple to reddish-green matrix supporting poorly-sorted angular lithic fragments up to 10 cm in size. Fragments consist of quartz, plagioclase and mafic and felsic volcanic rocks. Lahars are best exposed in the Duck Ponds area.

Volcanic breccias occur at the contact between the Frasers Brook Member and the overlying James River Formation. They consist of blocks of felsic to mafic volcanic rocks up to 5 m in diameter enclosed in a fine grained chloritic matrix. These rocks are exposed in the Keppoch Gulch area.

Finely laminated mudstones, by definition, occur only in the Moose River Member. They are black, grey and green, porcellaneous and are poorly cleaved. Graded bedding, load structures and CE Bouma sequences are commonly observed. These mudstones are virtually identical to those of the Morar Brook and Maple Ridge formations in the northern Antigonish Highlands (see below).

Petrography. Ignimbrites contain poorly-sorted angular clasts of porphyritic (plagioclase) rhyolite, quartz and plagioclase and subordinate orthoclase phenoclasts in a matrix dominated by flattened to ovoid pumice (Fig. 5). The angular shape of the clasts is commonly due to in situ brecciation. The pumice is commonly extremely elongate, defines the banding and imparts a eutaxitic primary foliation. It is extremely variable in abundance and consists of variably sericitized quartzofeldspathic material. Pumice fragments display wispy outlines, are moulded against crystal and lithic fragments and commonly merge imperceptibly with the matrix. The remainder of the matrix is dominated by cryptocrystalline quartzofeldspathic material displaying patchy recrystallization. Secondary minerals such as sericite and chlorite, or more rarely, calcite and epidote may also be present. Sericite and chlorite display a mimetic epitaxial relationship with the primary banding.

Lithic and crystal tuffs may be distinguished from ignimbrites by the lack of pumice and from flows by the clastic aspect of the matrix. The tuffs are moderately well sorted and massive. Lithic fragments are angular, are up to 30 mm in length and consist of plagioclase porphyritic rhyolites, similar to the lithic clasts in the ignimbrites. Broken angular crystal fragments of quartz and plagioclase also occur. The matrix is fine grained with a distinct clastic texture and is predominantly quartzofeldspathic in composition. It displays incipient to extensive alteration to sericite, epidote, and chlorite.

Felsic flows are generally aphyric but may contain quartz "eyes" and plagioclase phenocrysts up to 6 mm in length. The matrix consists of interlocking microphenocrysts of plagioclase and subordinate orthoclase displaying crude flow textures in a fine grained quartzofeldspathic matrix. The plagioclase is commonly saussuritized and the orthoclase may be sericitized. In some instances, the matrix displays spheroidal textures. Parts of the matrix also display a patchy recrystallization to randomly oriented coarse grains of quartz and feldspar. The matrix shows incipient to extensive alteration to epidote, chlorite, sericite, and biotite. Oxides are generally rare. Calcite is common adjacent to faults.

Basaltic andesites contain coarse phenocrysts of plagioclase up to 30 mm in length. In a few phenocrysts, relict oscillatory zoning is preserved, but the plagioclase is generally saussuritized to albite, epidote and minor calcite (Fig. 6). Rarely, augite microphenocrysts occur that are generally partially to completely replaced by chlorite. The matrix is dominated by albite (which displays crude flow alignment), epidote and actinolite, with relatively subordinate opaques, sphene, sericite, and quartz ± calcite.

Basalts consist of microphenocrysts of augite displaying an incipient alteration to chlorite, and abundant laths of plagioclase (albite) defining a good flow texture (Fig. 7). The basalts are readily distinguished from the basaltic andesites by the higher abundance of augite and opaques and by the paucity or absence of actinolite and quartz. Opaques are titaniferous magnetite or more rarely pyrite, and are equant to interstitial. Secondary minerals consist of epidote and chlorite.

Lahars are extremely heterogeneous in texture and composition (Fig. 8). These rocks are both matrix and clast supported. Fragments are angular to subrounded and are poorly sorted. The fragments consist of felsic and mafic volcanic rocks that strongly resemble those of the Keppoch Formation described above, except that they display moderate to strong hematitic alteration. The matrix consists of abundant hematite, chlorite, and epidote with variable proportions of quartz, sericite, biotite, actinolite, and carbonate.

The mudstones consist of fine grained subangular to subrounded quartz, albite, and minor pyrite in a matrix dominated by sericite, chlorite, and minor biotite, which define a weak cleavage generally at a shallow to moderate angle to bedding. Coarser grained units display graded bedding.

Environment of deposition. The lack of marine sediments, the weathered tops of volcanic units and the presence of welded ignimbrites indicate that the Frasers Brook Member was deposited in a subaerial environment. Thick proximal sequences of felsic or mafic volcanic rocks occur in the core of the Keppoch block whereas interlayering of the two is typical at the extremities of the Keppoch block. This suggests a central topographic high surrounded by lower areas where both proximal and distal lavas intermix.

The Moose River Member was probably deposited in a submarine environment because the mudstones and greywackes are inferred to have been deposited by turbidity currents. The fine planar lamination suggest a low energy environment of deposition, possibly at the distal end of a submarine fan model (see Walker, 1984). Individual mudstone units commonly attain a thickness of 1 to 6 m.

Figure 5. Photomicrograph (4 mm wide) of an ignimbrite, Keppoch Formation. Note the pumice in the upper part of the photograph. The matrix defines a locally irregular foliation, is cryptocrystalline and is incipiently altered to sericite.

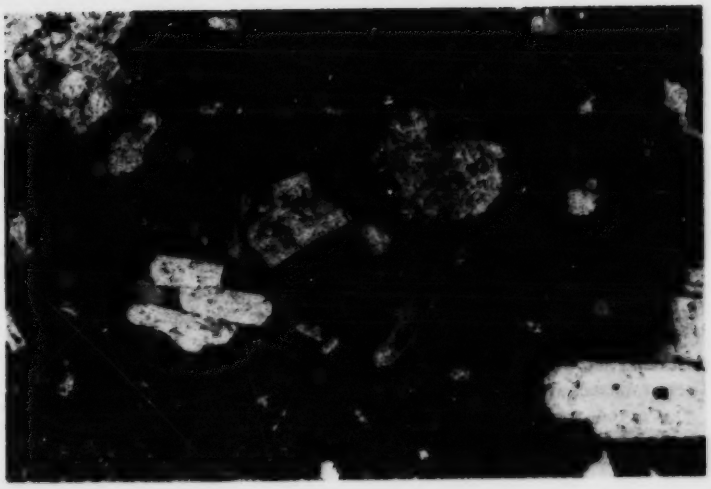


Figure 6. Photomicrograph (4 mm wide) of a basaltic andesite, Keppoch Formation. These rocks are characterized by microphenocrysts of plagioclase (now albite) in a fine grained groundmass of plagioclase, chlorite, actinolite, epidote and opaques.

Chisholm Brook Formation

Field relationships. The Chisholm Brook Formation outcrops only in the Georgeville block (Fig. 3). Its type section occurs along the shoreline near the mouth of Chisholm Brook (Fig. 2 and 4). Other good sections occur in Chisholm Brook and along the shoreline west of Morar Brook. The base of the formation is not exposed. However, the top of the formation is exposed on the shoreline near the northern margin of the Greendale Complex and at Browns Brook. At the former locality, interbedded volcaniclastic rocks and mudstones grade stratigraphically upward over 20 m (the contact is overturned) into mudstones and cherts. The top of the Chisholm Brook Formation is placed at the top of the uppermost volcaniclastic unit clearly recognizable in the field. At Browns Brook, an 8 m thick mafic dyke has intruded along the contact between the volcanic-volcaniclastic rocks and the mudstone-chert-limestone assemblage, obscuring the original stratigraphic relationships.

The lack of exposure of the base of the formation and the scarcity of way-up criteria, together with the presence of recumbent folds, make it impossible to determine the thickness of the formation. The type section exposes about 600 m of uniformly dipping rocks; however, structural repetitions probably exist in this section.

Field description. The Chisholm Brook Formation consists of metamorphosed green fine grained aphyric mafic lavas and pyroclastic rocks with minor interbedded white-grey banded marble and dark grey mudstone. Flow boundaries were observed only where the volcanic rocks are interbedded with

Figure 7. Photomicrograph (4 mm wide) of a basalt, Keppoch Formation, which contains laths of plagioclase, opaques (titanomagnetite) and relict augite (upper right). See text for further description.

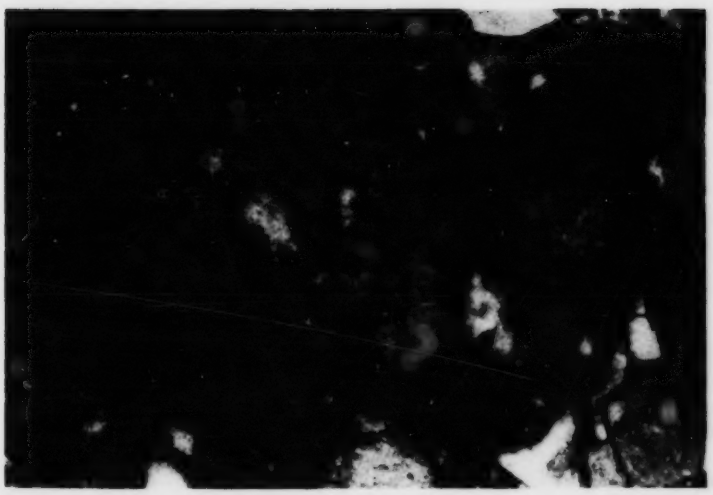


Figure 8. Photomicrograph (4 mm wide) of a lahar. The lahar contains angular volcanic fragments of various lithologies typical of the Keppoch Formation. Note the fragment of ignimbrite in the upper central part of the photograph. The matrix is dominated by chlorite and hematite.

marbles. Individual flows range up to 4 m in thickness. Pyroclastic rocks vary from lapilli tuff to volcanic breccia with fragments as large as 60 cm in length. They are generally massive and bedding is often obscure; however, graded bedding is observed locally. Adjacent to the Greendale Complex, the Chisholm Brook Formation has been thermally metamorphosed and altered (Murphy, 1982). Within the complex, some xenoliths of basalt and marble appear to retain their structural continuity with the host rocks and many are probably roof pendants. Marble horizons within the Greendale Complex may be as thick as 4 m and vary from pure white marbles to green impure marbles.

On the shoreline west of Morar Brook, the Chisholm Brook Formation consists of fine- to medium-grained tuff and mafic lava, and is cut by numerous mafic dykes and sheets. Marbles and mudstones are absent. These observations, the circular outcrop pattern and the dyke along the contact with the Morar Brook Formation suggest that this area may lie close to, or above, a vent complex. Inland exposures of the Chisholm Brook Formation are limited and are mainly of mafic volcanic rocks, which are more resistant to erosion than the marbles.

Petrography. In thin section, primary textures are preserved in the basalts but the mineralogy is entirely secondary. The mineralogy is dominated by albite, actinolite, epidote, chlorite, quartz and sphene ± calcite. This assemblage is typical of the greenschist facies of metamorphism. Garnet and hornblende are common in the basalts adjacent to the Greendale Complex, and are attributed to contact metamorphism which clearly overprints the regional metamorphism.

The rocks are dominated by slender laths of albite and actinolite which generally display a rough parallelism. Rarely, plagioclase (now albite) microphenocrysts up to 2 mm in length occur. They have an extremely ragged shape and are partially replaced by epidote, actinolite, and chlorite. Epidote forms minute anhedral blebs and is homogeneously distributed throughout the matrix. Actinolite and calcite have an antipathetic relationship. Chlorite shows textures similar to those of actinolite, whereas calcite forms ragged grains up to 2 mm in length. Quartz is typically interstitial. Prehnite occurs exclusively in veins.

Dykes are spatially associated with these volcanic rocks, especially in the Browns Brook area. Centres of dykes are relatively unaltered and contain relict microphenocrysts of augite (partially altered to actinolite) and plagioclase (now albite) in a subophitic intergrowth, in a matrix of albite, chlorite, actinolite, and quartz. One dyke contains large feldspar phenocrysts up to 10 mm in length which show relict zoning. The cores of these crystals contain abundant epidote inclusions whereas the rims are relatively fresh and are albitic in composition.

Marbles are saccharoidal in texture, predominantly consisting of calcite, tremolite, serpentine (after forsterite) ± chlorite ± epidote, and quartz. Calcite, the most abundant phase, forms minute ragged interlocking grains. Tremolite occurs as thin laths and is commonly partially altered to chlorite. Serpentine is generally present, forming small globular blebs and pseudomorphs after calcite. Rarely, forsterite is preserved in the centres of crystals, though is not in equilibrium with the other phases in the rock. Epidote and quartz have variable abundances, but together they generally constitute less than 5 percent of the rock by volume. A comparison of the mineralogy of the marble with that of the calcareous rocks in the Morar Brook Formation (discussed below) indicates that the assemblages here are due to contact metamorphism of the rocks adjacent to the intrusive Greendale Complex.

Envrionment of Deposition. The geochemistry of the Chisholm Brook mafic volcanic rocks (see the section on Geochemistry below) suggests that they were extruded during the incipient stages of rifting of a small basin associated with a volcanic arc. This is consistent with the shallow water environment suggested by the interbedded marble.

Morar Brook Formation

Field relationships. The Morar Brook Formation is restricted to the Georgeville block (Fig. 3 and 4), although it is correlated with the Maple Ridge Formation farther south. The type section of the Morar and Horseshoe Turn brooks (Fig. 2). The Morar Brook Formation has conformable contacts with both the Chisholm Brook Formation below and the Livingstone Cove Formation above. Recumbent folding in the Morar Brook Formation precludes any estimate of true thickness; however, it has a tectonic width up to 700 m.

Field description. The Morar Brook Formation consists mainly of black, grey, and green porcellaneous iron-rich finely laminated poorly cleaved fine grained mudstone and siltstone, and some thin beds of chert and limestone. Graded bedding, cross-bedding and load casts are common. Slump folds are locally common though sporadically distributed. These folds have irregular styles and orientations and are generally found within thin mudstone units that are interlayered with thicker siltstones. The folds are generally tight to isoclinal and are commonly reclined. ACE and CE Bouma sequences were observed, some of which have current-reworked tops. An analysis of the primary structures indicates that the beds face both up and down in uniformly dipping sections, indicating the presence of large tight-isoclinal intrafolial folds.

Petrography. The black laminated mudstones of the Morar Brook Formation consist of fine grained subangular to subrounded quartz and albite with rare rutile in a matrix that is dominated by sericite, chlorite, and minor biotite. Pyrite is generally more abundant in the darker laminae and occurs in elongated aggregates. Although graded bedding may be difficult to observe in the field, it is commonly visible in thin section in some of the coarser grained units. Matrix minerals display a strong preferred orientation consistently parallel to a fracture cleavage. In most cases, this orientation is also parallel to bedding. Crosscutting relationships between

cleavage and bedding are rarely visible in the field. However, microscopic inspection of some of the hinge zones of minor folds shows that this relationship exists here (Fig. 9).

The siltstones have similar petrographical characteristics, except that they are coarser grained and contain variable proportions of lithic fragments. The fragments generally consist of slate or epidote-bearing volcanic rocks. The cherts are very pure, consisting of interlocking grains of microcrystalline quartz showing undulose extinction. The cherts may contain minor disseminated grains of pyrite.

Environment of deposition. The mudstones and siltstones, which make up the bulk of the Morar Brook Formation, are inferred to have been deposited by turbidity currents. The common occurrence of slump structures suggests a proximal base-of-slope environment, perhaps as levees or back levees in a submarine fan (Walker, 1984). However, the thin beds of chert and limestone suggest periods of quiet, deep water sedimentation as well.

Livingstone Cove Formation

Field relationships. The Livingstone Cove Formation occurs only in the Georgeville block (Fig. 3 and 4). Its type section occurs along the shoreline southwest of Livingstone Cove (Fig. 2) where it rests conformably on the Morar Brook Formation (Fig. 3). The base of the Livingstone Cove Formation is defined as the first appearance of green conglomerate. This contact may be observed at several localities along the shoreline.

Field description. The lowest conglomerate, in common with most other conglomerates higher in the Livingstone Cove Formation, has a channelled base truncating the bedding in the underlying mudstones. The black mudstones and thin cherts interbedded with the conglomerates of the Livingstone Cove Formation are identical to those of the underlying Morar Brook Formation, suggesting that the stratigraphic record is essentially continuous. At Livingstone Cove, the Livingstone Cove Formation is unconformably overlain by red conglomerates and sandstones of the Middle Devonian McAras Brook Formation. At this locality, the Livingstone Cove Formation is overturned and dips gently northeastward. The McAras Brook Formation is the right way up, gently dips to the north and rests with angular unconformity upon the mudstones and siltstones of the Livingstone Cove Formation. The unconformity is displaced by a series of minor northeast-trending faults.

It is difficult to determine the thickness of the Livingstone Cove Formation. Along the shorelines it has a minimum tectonic width of 700 m, but at least three fold repetitions occur in this section and thrusting has been observed. The stratigraphic top to the Livingstone Cove Formation was not observed.

The Livingstone Cove Formation is made up of green conglomerate and greywacke interbedded with black mudstones and minor thin cherts and limestones. The conglomerates become thicker and more common higher up in the stratigraphic section, and are generally matrix supported. They vary from unsorted beds with little or no internal structure, through beds displaying both inverse and normal grading and imbrication, to pebbly sandstones with normal grading and cross-stratification. The clasts are usually rounded, spherical or elliptical, are up to 10 mm in size and are supported by a sandy volcaniclastic matrix. Pebbles in the conglomerates are predominantly of volcanic origin and comprise intermediate to felsic lava, crystal tuff and ash, granite, jasper, rose quartz, quartz and feldspar with locally derived black mudstone and siltstone.

Petrography. With the exception of the conglomerates and associated sandstones, the rock types in the Livingstone Cove Formation are the same as those in the Morar Brook Formation; thus, only the conglomerates will be described here. Dacitic to andesitic conglomerate pebbles are devoid of phenocrysts and consist almost entirely of thin laths of albite

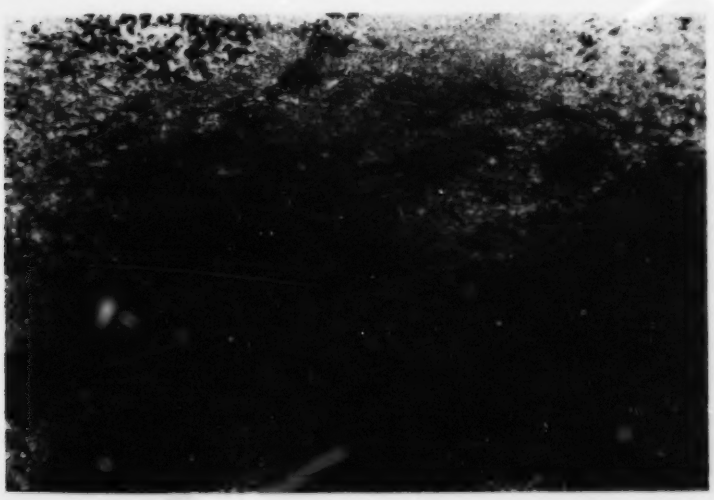


Figure 9. Photomicrograph (4 mm wide) showing S_0/S_1 relationships in the hinge of an F_1 minor fold in the Morar Brook Formation. Bedding is defined by siltstone-mudstone bands. Note that the siltstones contain opaques (magnetite and pyrite).

displaying trachytic textures, with minor quartz and opaques. Rhyolitic pebbles contain phenocrysts of orthoclase, albite and chlorite enclosed in a fine grained matrix of albite, chlorite, epidote, and quartz. Pebbles of tuff contain microphenocrysts of albite, or quartz (or both) enclosed in a cryptocrystalline matrix. Some tuff clasts may be ignimbrite because they contain flattened pumice. Granite clasts commonly display graphic textures and contain quartz, plagioclase and potassium feldspar. Siltstones and black mudstones compose about half of the pebbles and have petrographical characteristics similar to those of the Morar Brook Formation. Clastic crystal fragments of albite and quartz are generally angular to subangular. The matrix of the conglomerates is largely clastic, consisting of angular fragments of quartz, albite, siltstone, and black mudstone. However, secondary minerals such as chlorite, sericite, and epidote are also common.

Greywackes are finer grained, relatively well sorted rocks. They are more quartz rich than the conglomerates but contain similar clastic lithologies though in lesser abundance.

Environment of deposition. The conglomerates interbedded with the mudstones and siltstones in the Livingstone Cove Formation are inferred to have been deposited in submarine feeder channels near the base of the slope. The unsorted conglomerates are inferred to be debris flows deposited at a change in slope, and the graded conglomerates and greywackes were probably deposited a little farther down the channel (Walker, 1984). A channel deposit origin for the conglomerates may account for the lenticular shape of some of the outcrop patterns (although folding undoubtedly complicates the picture) and for the channelled bases. The interbedded slumped mudstones and siltstones, which are similar to those in the Morar Brook Formation, are inferred to have been deposited in a proximal levee or back levee environment. The upward thickening and coarsening sequence suggests that the fan was prograding. The rounded nature of the pebbles in the conglomerates suggests that they were derived from shallow water gravels on a shelf. The intermediate to felsic volcano-plutonic nature of the pebbles indicates that this shelf bounded volcanic highlands of a volcanic arc. The siltstone and mudstone clasts were probably derived from the sides of the submarine canyon. The fine grained clastic sediments interbedded with the conglomerate may have been derived from the outer shelf environment at the top of the slope.

Maple Ridge Formation

Field relationships. The Maple Ridge Formation occurs only in the Maple Ridge block (Fig. 3). The type section occurs in East Doctors Brook, and the formation is well exposed in the Eigg Mountain and Bears Brook area (Fig. 2). The base of the formation is not exposed, but the top of the formation, while not exposed, is closely defined by outcrops on the northernmost tributary of Rights River. The upper contact of this formation with the lower member of the James River Formation is defined by the first appearance of matrix-supported conglomerate. Where the James River Formation contains no conglomerate, the upper contact of the Maple Ridge Formation is defined by the appearance of thick greywacke sequences.

Field description. Typically, the formation consists of finely laminated iron-rich porcellaneous black and green poorly cleaved mudstones and siltstones (~55%) and thinly bedded green greywackes (~40%) in which individual beds are less than 2 m thick. Magnetite-rich mudstone forms a useful marker horizon that can be traced on aeromagnetic maps (Geological Survey of Canada, 1982). Green mafic flows form a minor part of the formation. Poor exposure combined with isoclinal folding preclude any reliable estimate of the thickness of the formation, but it is about 3 km wide. Primary structures include graded bedding, cross-bedding and slump structures. CDE, ACDE and ABCDE Bouma sequences were observed.

Petrography. The mudstones and siltstones of the Maple Ridge Formation consist of silt- to clay-sized angular fragments of quartz and albite ± pyrite dispersed in a matrix of sericite and chlorite, with calcite present in some of the lighter coloured laminae. It is not certain whether the albite and pyrite are detrital, authigenic or secondary in origin. One of the characteristics of the mudstones of this formation is the blocky nature of the rock where cleavage is parallel to bedding. Graded bedding is especially obvious in this section.

The wackes consist of angular to subangular clasts of lithic fragments (~50%), quartz (~40%) and albite (10%) in a matrix of quartz, chlorite, and epidote. The lithic fragments consist almost entirely of intermediate to felsic volcanic rocks. The wackes are classified as litharenites or sublitharenites.

The petrography of the mafic rocks is similar to that of the Clydesdale Formation, which is described below.

Environment of deposition. The interbedded mudstones, siltstones, and wackes of the Maple Ridge Formation appear to alternate between thinly bedded slumped turbidites and classical turbidites. In a submarine fan model (Walker, 1984) this sequence is explicable in terms of either: (i) a proximal levee or back levee environment onto which a turbiditic fan periodically encroached; or (ii) a mid to lower fan environment bounded by a syndepositional fault scarp adjacent to which slumping took place. A volcanic arc source is indicated for the lithic clasts.

James River Formation

Field relationships. The James River Formation outcrops in the Maple Ridge, Clydesdale, and Keppoch blocks (Fig. 3). The type section occurs in James River immediately southeast of the Browns Mountain Fault (Fig. 2 and 3). Other good sections occur in South Rights River and in the Mayfield area. In the Maple Ridge and Clydesdale blocks, the formation is divided into two members. The lower member is defined by the presence of matrix-supported conglomerates and is

bounded by the base and the top of the lowest and highest conglomerates. These conglomerates are interbedded with green shales, siltstones, and wackes. The upper member is made up of wackes, siltstones and shales only, with finer lithologies becoming predominant upward. Matrix-supported conglomerates are rare in the Keppoch block and the division into two members could not be made.

The base of the James River Formation is not exposed in the Maple Ridge and Clydesdale blocks. Here, the boundary between the Maple Ridge and James River formations may only be defined in a series of isolated outcrops in the upper reaches of the northernmost tributary of Rights River, where it is positioned between the thinly laminated mudstones (Maple Ridge Formation) apparently conformably underlying conglomerates and greywackes (James River Formation, lower member). The conformable contact between the lower and upper members is exposed in both the type section and the upper reaches of South Rights River. The upper contact of the James River Formation with the Clydesdale Formation is gradational and is placed where mafic lava flows appear in the succession on South Rights River. The lower member is at least 400 m thick and the upper member is about 250 m thick, for a total of at least 650 m.

In the Keppoch block, the contact between the James River and Keppoch formations is conformable in the west but is unconformable in the east. The lower contact of the James River Formation is defined where volcanic rocks are minor. In the west, this contact is gradational from subaerial volcanic rocks (Keppoch Formation, Frasers Brook Member) through submarine volcanic rocks and finely laminated mudstones (Keppoch Formation, Moose River Member) to greywackes and minor mudstones (James River Formation). In the east, the James River Formation rests directly on the Frasers Brook Member with the Moose River Member missing. The thickness of the James River Formation in the Keppoch block is about 2000 m.

Field description. The conglomerates are very similar to those of the Livingstone Cove Formation in lithology and clast type. However, the pebbles are generally more rounded and smaller (up to 1.5 cm). They are generally normally graded and are rarely unsorted or inversely graded. The shales, siltstones and greywackes are similar to those in the underlying Maple Ridge Formation, except that (i) slump structures are rare, (ii) they are not as thinly bedded, (iii) the cleavage is not as well developed and (iv) they tend to be more siliceous. The wackes are typically 1 to 4 m thick. In view of these similarities, no further lithological or petrographical description is needed here.

Environment of deposition. The conglomerates of the James River Formation are inferred to occur in channels feeding a turbidite fan. The smaller size of the clasts and the predominance of normal grading suggest that these channels were farther away from the base of the slope than those in the Livingstone Cove Formation. An inner fan or suprafan channel environment is inferred (cf. Walker, 1984). The interbedded wackes, siltstones and shales indicate a mid-fan environment. The thinning and fining upward sequence observed in the James River Formation suggests a gradual regression of the fan, perhaps due to progressive change in the locus feeder channel.

Clydesdale Formation

Field relationships. The Clydesdale Formation occurs in the Clydesdale block and the southern part of the Maple Ridge block (Fig. 3). The type section occurs in South Rights River west of Clydesdale (Fig. 2). Other good sections occur farther upstream in South Rights River, in James River and in several southwesterly flowing tributaries of Brierly Brook. The base and top of the formation are exposed in South Rights River and in James River, and are defined by the first and last appearance of mafic flows interbedded with green greywackes and shales. The mafic flows are the predominant lithology in this formation. The formation is approximately 400 m thick on South Rights River.

Field description. The volcanic units are green, aphyric, fine grained and spatially associated with mafic dykes, which are difficult to distinguish from flows when contacts are not exposed. The wackes and shales are similar to those of the upper member of the James River Formation. A north-south line of positive residual total field and vertical gradient magnetic anomalies occurs over the central part of the Clydesdale Formation (Fig. 10, after Geological Survey of Canada, 1982). Unfortunately, outcrop exposure is lacking along these anomalies, although adjacent exposures commonly contain mafic dykes. It is possible that the anomalies represent the intrusive rocks that acted as feeders for the mafic Clydesdale volcanic rocks. The Clydesdale Formation dies out to the south where the James River Formation is conformably overlain by the South Rights Formation (Murphy, 1984). The magnetic anomalies appear to be restricted to the extent of the Clydesdale Formation.

Petrography. The mineralogy of the Clydesdale basalts is predominantly secondary with relict primary phases; however, the primary igneous textures are often well preserved. Typically, basalts contain abundant thin laths and rare microphenocrysts of feldspar (now albite) and exhibit good flow textures. Microphenocrysts of augite may be present as a relict primary phase. Augite displays subophitic relationships with plagioclase, but is generally partially or completely altered to chlorite and, more rarely, to actinolite. Chlorite is also abundant as a secondary phase in the matrix, where it is commonly associated with epidote and pumpellyite. Magnetite and rutile are minor accessory minerals. Calcite is rare and tends to occur in vesicles, where it is rimmed by opaques which are in turn enclosed by chlorite.

The petrography of the interbedded wackes and shales is similar to that of the Maple Ridge Formation and will not be repeated here.

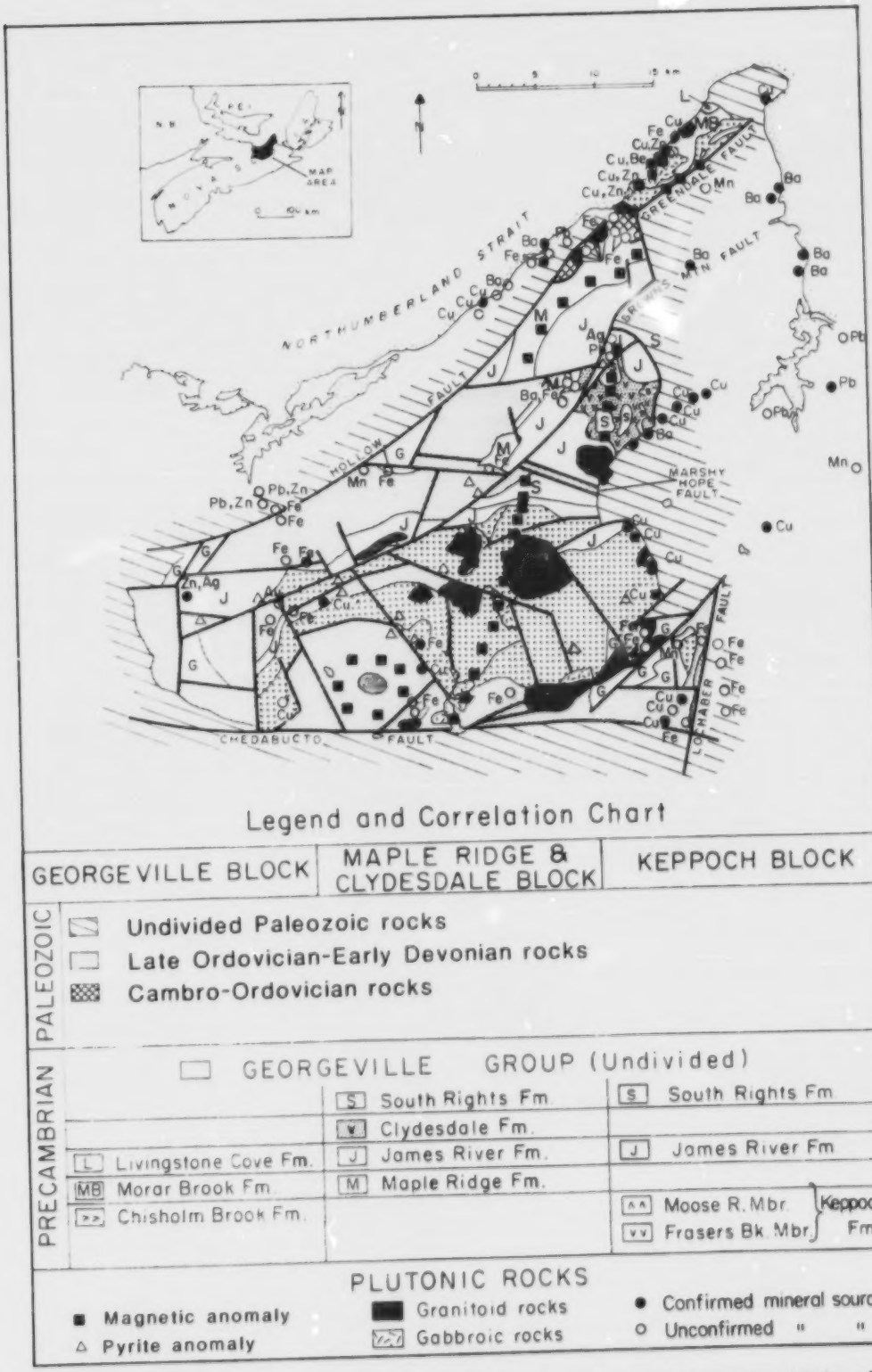


Figure 10. Summary geological map showing the locations of mineral occurrences (circles after Bourque, 1981; triangles, this study) and vertical gradient magnetic anomalies (see Geological Survey of Canada, 1982 for further details).

Environment of deposition. The geochemistry of the Clydesdale mafic volcanic rocks suggests an ocean island tectonic setting (see the Geochemistry section below). This suggests that the interbedded sedimentary rocks were deposited in a marine basin floored by attenuated continental crust. The line of magnetic anomalies in the area associated with the Clydesdale Formation may be representative of vent intrusive rocks. Their linear arrangement might conceivably be related to hot spot activity beneath a moving plate, producing a gradual migration of volcanic activity with time. Alternatively, it could be a fault zone along which magma was preferentially emplaced.

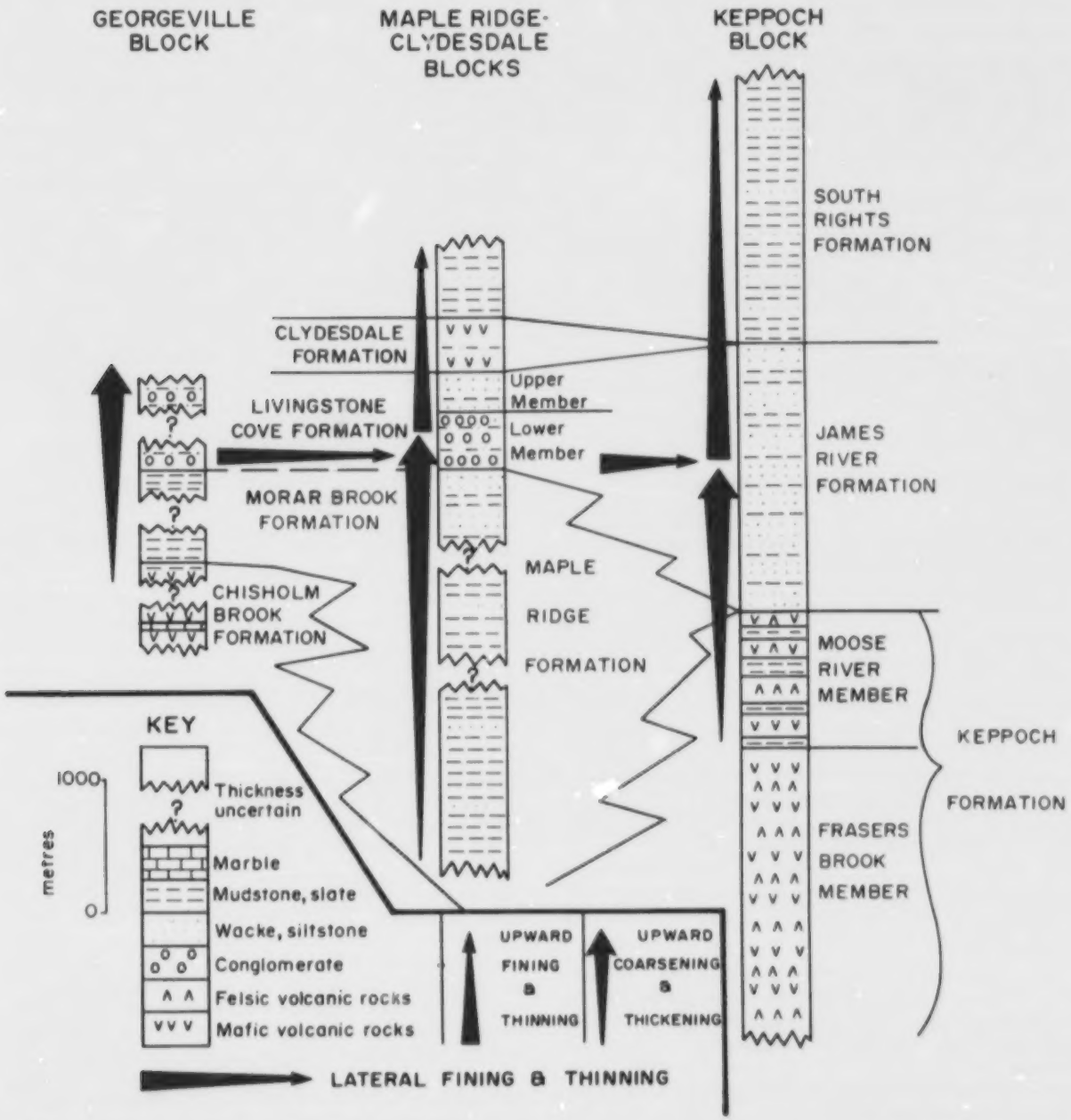


Figure 11. Stratigraphic correlation chart for the Antigonish Highlands. Thicknesses of the Chisholm Brook, Morar Brook, Livingstone Cove and Maple Ridge formations are unknown due to fold and fault repetitions.

South Rights Formation

Field relationships. The South Rights Formation outcrops in the Clydesdale block, in the central part of the Maple Ridge block and in the northern part of the Keppoch block (Fig. 3). Its type section occurs in the upper reaches of South Rights River with another good section occurring farther downstream (Fig. 2). The formation consists almost entirely of green finely laminated shales and siltstones which are commonly graded and cross-bedded and display ABCE, BCE and CE Bouma units. Slump structures are absent. Rare outcrops of wacke and mafic volcanic rocks also occur. The base of the formation is placed at the top of the mafic flows. The top of the unit is not exposed. The unit is at least 400 m thick in the type section and is about 200 m thick in the Keppoch block. In the field and in thin section, the sediments are similar to those of the Maple Ridge Formation.

Environment of deposition. The dominance of planar laminated mudstones suggests a lower energy environment of deposition than that of the underlying James River Formation. The thinly bedded character of these turbidites and the general absence of slump structures suggests either a proximal levee, back levee or a distal outer fan environment. The gradational contact of this formation with the underlying James River Formation in the Keppoch block (which probably has an inner fan environment) suggests that the fan was regressive.

Correlations within the Antigonish Highlands

The correlation between the units in the various fault blocks is shown in Figure 11. The correlation between the Georgeville, Maple Ridge and Clydesdale blocks is based upon: (i) the correlation of the conglomerates in the Livingstone Cove Formation with the lower member of the James River Formation; (ii) the similarity between the Morar Brook and Maple Ridge formations; and (iii) the coarsening upward sequences that are common to the lower parts of all three blocks. Although the sequence observed in the Georgeville block is similar to that in the Maple Ridge and Clydesdale blocks, the following differences occur: (i) wackes, which form an integral part of most of the units in the south are rare in the Georgeville block; and (ii) the conglomerates in the Georgeville block are coarser, thicker and less sorted than those occurring in the south. These differences may be explicable in terms of facies changes. It is also possible, however, that a turbidite feeder channel represented by the conglomerates could have occurred at different times in a small ocean basin.

The geochemistry of the basaltic andesites of the Chisholm Brook and Keppoch formations (see the section on Geochemistry below) are remarkably similar and suggests that these formations may be correlated. Both these sequences are unusual in that they display a similar transitional chemistry between tholeiitic and calc-alkalic suites (see the Geochemistry section below). In the Keppoch block, the James River Formation directly overlies the Keppoch Formation. The above correlations imply that the lower part of the Keppoch Formation is equivalent to the Chisholm Brook Formation while the Moose River Member of the Keppoch Formation is a facies variation of the Maple Ridge and Morar Brook formations.

Depositional model for the Georgeville Group

If the above correlations are accepted then it is possible to speculate on a facies model for the Georgeville Group. The stratigraphy records the development of a sedimentary basin (Fig. 12). The overall distribution of the formations (although modified by deformation) supports a basin model with the lowest formations occurring in the northern and southernmost blocks and the highest formations occurring in the central blocks (Fig. 3). The lowest units (the Chisholm Brook and Keppoch formations) were deposited in a shallow marine or subaerial environment, and are overlain by a thick sequence of turbidites. The environment of deposition of the turbidites can be accommodated in Walker's (1984) submarine fan model. In the Georgeville block, the sequence thickens and coarsens upward; in the Maple Ridge and Clydesdale blocks, the succession at first thickens and coarsens upward, then is followed by a thinning and fining upward sequence. Also, a lateral change from north to south shows: (i) a fining of the conglomerates; (ii) a decrease in the frequency of slump structures; and (iii) the inferred proximal upper channel back levee environment passes into a more basinward mid-fan environment.

A possible model for the fine-coarse-fine vertical sequence observed in the south is a submarine canyon which moved laterally through time by transcurrent faulting (Fig. 13). Thus, the sequence in a given area would coarsen as the channel approached and would become finer as the channel receded. The presence of transcurrent faults during deposition is supported by the limited temporal gap between deposition deformation and intrusion of the Georgeville Group. Alternatively, the submarine canyon may have changed its location as a result of other factors. Farther south the Keppoch block became submerged at some stage and was draped by mid-fan to distal turbidites (Fig. 12).

A basin model is supported by the geochemistry of the volcanic rocks (see the Geochemistry section below). Thus, the Chisholm Brook Formation is believed to represent the initial rifting stage in the formation of the Georgeville Basin. The Keppoch Formation is transitional between a calc-alkaline volcanic arc and a rifting environment, whereas the Clydesdale Formation is interpreted in terms of tholeiitic to alkalic magmatism erupted into attenuated continental crust (Fig. 4). Keppie (1982) proposed that the Georgeville Basin formed as an interarc basin within a volcanic arc which was split to form the Jeffers (Cobequid Terrane) and Fourchu (Cape Breton Terrane) volcanic arcs. However, the intracontinental ntaure of the Clydesdale volcanism and the similarities between Antigonish and Cobequid straigraphy suggest that the marine basin was not floored by oceanic crust. Comparisons with modern volcanic arc complexes suggest that basins can often form as pull-apart basins in association with transcurrent faulting along the volcanic arc axis as a result of oblique convergence (Fitch, 1972).

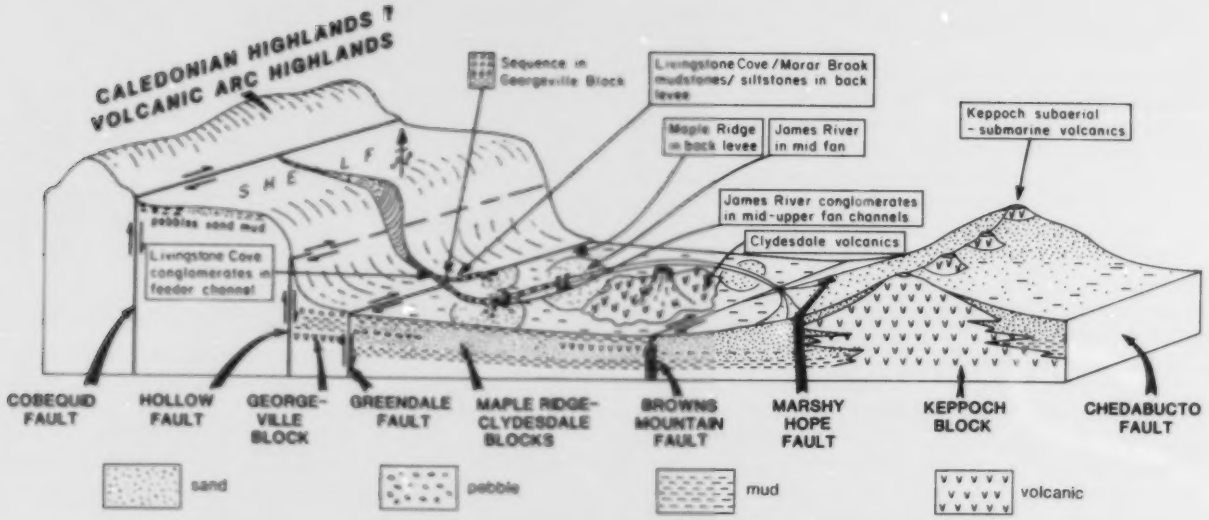
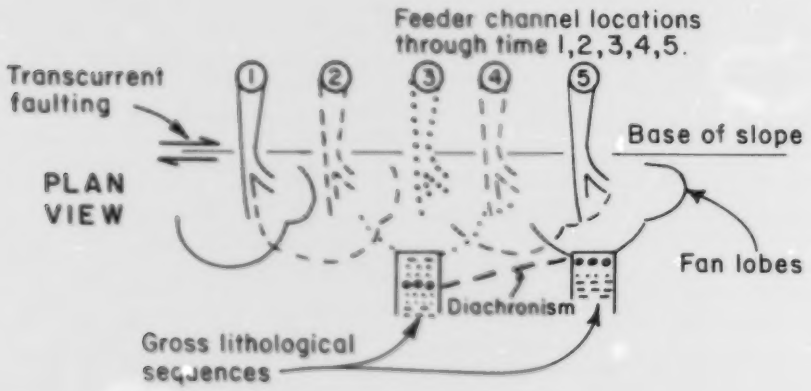


Figure 12. Depositional model for the Georgeville Group. Location of fault-bounded blocks shown in Figure 3.



Transcurrent faulting beheads the feeder channel

The turbidite fan moves laterally through time resulting in the following succession of environments:

MARGIN →		BASIN CENTRE
Back levee	Lower fan	Fine - thin
	Mid fan	↑
Upper fan with channels	Supra fan with channels	Coarse - thick
	Mid fan	↑
Back levee	Lower fan	Fine - thin

Figure 13. Stratigraphic sequence produced by a turbidite fan progressively displaced by transcurrent faulting.

The source area of these turbiditic sediments is uncertain due to the paucity of current direction data and the effects of polyphase deformation. The composition of the pebbles and the presence of 613 ± 5 Ma detrital zircons (Keppie and Krogh, 1990) indicates that the source area was probably a penecontemporaneous, emergent volcanic arc. Thus the Georgeville Group sediments may be classified as synorogenic. Possible candidates for late Precambrian volcanic arc complexes occur in the Cobequid Highlands (Jeffers Formation) to the northwest (Donohoe and Wallace, 1980; Pe-Piper and Piper, 1987), in the Keppoch Formation of the southern Antigonish Highlands (Murphy, 1984) and in southeastern Cape Breton (Fourchu Group, Keppie et al., 1979). Sedimentological relationships described above indicate that the Cobequid Highlands are the most likely source area. It is envisaged that an eroding volcanic arc source occurred to the northwest in the Cobequid Highlands, and was fringed by a narrow fault-bounded shelf. This is consistent with the presence of 613 ± 5 Ma detrital zircons within the Livingstone Cove Formation. This shelf was dissected by submarine canyons along which turbidity currents flowed periodically carrying volcanic pebble detritus, sand and mud. These were carried southeastward into the basin and deposited as submarine turbidite fans. In this configuration, the Georgeville block is proximal and the Maple Ridge-Clydesdale blocks lie basinward.

Geochemistry

Analytical notes

Ninety-eight samples of volcanic rocks from the Keppoch, Chisholm Brook and Clydesdale formations were analyzed for major and trace (Ba, Rb, Sr, Zr, Nb, Y, Ga, Ni, V, Cu and Zn) elements by X-ray fluorescence at X-Ray Assay Laboratories, Toronto, and at St. Mary's University, Halifax. Thirty-three duplicate samples were also analyzed. The precision and accuracy of these determinations are generally about 5 percent, with the exception of the Nb and Cr analyses (generally between 5 and 10 percent). The rare earth elements were analyzed by instrument neutron activation at Neutron Activation Services, McMaster University, with a precision between 5 and 10 percent. Pyroxenes were analyzed by energy dispersive microprobe at Dalhousie University. The full sets of analyses displaying the geochemistry of the Chisholm Brook, Clydesdale and Keppoch formations are shown in Tables 1, 2 and 3, respectively. Petrographical evidence of alteration in all of these formations is reflected in the wide ranges in the concentrations of Na_2O, K_2O, CaO, Rb and Sr (Tables 1, 2 and 3). For example, in the Chisholm Brook Formation, Na_2O varies from 2.1 to 4.6 weight percent, K_2O from 0.3 to 2.1 weight percent and Rb from 6 to 58 ppm for the mafic flows. For all of the formations, variation diagrams involving these elements give erratic trends (not shown). On the other hand, many elements (e.g. Si, Al, Mg, Fe, Ti, P, Zr, Y, Nb and REE) show more restricted ranges in absolute abundances and interelement ratios, and display typical magmatic trends.

SiO_2 content ranges from 49 to 62 weight percent in the flows of the Chisholm Brook Formation and from 43 to 55 weight percent in the Clydesdale Formation (on a volatile-free basis). Although some of this range may be due to secondary effects, the SiO_2 content confirms the basaltic to intermediate character of Georgeville volcanism in the northern Antigonish Highlands. In contrast, the Keppoch Formation in the southern Antigonish Highlands is bimodal with respect to SiO_2 (Fig. 14) with only 5 analyses (out of 66) containing between 55 and 65 weight percent SiO_2 (on a volatile-free basis).

Geochemistry of the Chisholm Brook Formation

Magmatic affinity. The basaltic rocks of this formation may be subdivided into low Ti (TiO_2 <1.4 weight percent) and high Ti (TiO_2 >1.4 weight percent) groups. Although the Zr/P_2O_5 versus TiO_2 discrimination diagram is inconclusive (Fig. 15a), the SiO_2 versus Zr/TiO_2 and Nb/Y discrimination diagrams (Fig. 15b,c) show that the low Ti rocks are predominantly subalkaline basalts with minor andesites. The low Zr/TiO_2 in these rocks is typical of either tholeiitic or calc-alkalic suites. The low Ti rocks show a small variation in FeO^t/MgO and a significant range in SiO_2 that is typical of calc-alkalic suites (Miyashiro, 1974; Fig. 16a), although the low (Mg) values (i.e. low Mg/Mg+Fe, with $Fe^{3+}/Fe^{2+} = 0.15$) and high concentration of FeO^t and FeO^t/MgO indicate that these rocks had a significant iron enrichment at an early stage in their evolution (Fig. 16b). The abundances of transition elements including Cr and Ni are low and are typical of calc-alkalic basalts that have undergone extensive fractionation. The chondrite-normalized REE abundances display a weak LREE enrichment (Fig. 17a) with La/Sm ranging from 2.5 to 4 and La/Yb ranging from 3 to 8, and closely resemble the REE patterns of calc-alkalic basalts (Pearce and Gale, 1977; Gill, 1981).

The high Ti basalts are similar to the basalts of the Clydesdale Formation and the aphyric basalts of the Keppoch Formation. Their geochemistry is typical of within-plate continental tholeiitic or alkalic rocks (see below).

Tectonic setting. The MORB-normalized multielement plot for the low Ti rocks (Fig. 18) shows a moderate enrichment in most incompatible elements relative to N-type MORB, and a depletion of Ni, Cr and Y. Such profiles, which also display

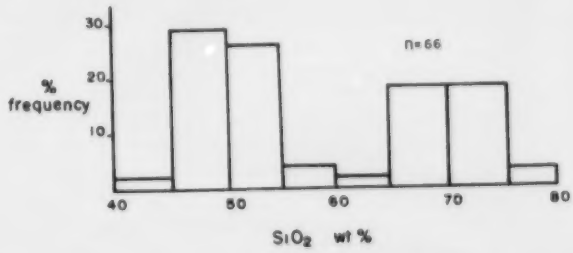


Figure 14. Frequency histogram of the SiO_2 content of volcanic rocks of the Keppoch Formation.

Table 1. Geochemistry of the Chisolm Brook Formation

	Mafic Flows											
	E16-207	E16-556	E16-714	E16-554	E16-206	E16-217	E16-215	E16-209	E16-214	E16-721	E16-547	E16-724
SiO_2	48.2	48.5	48.8	50.2	51.4	51.5	52.2	52.6	53.0	53.1	53.2	54.1
TiO_2	2.0	1.6	1.4	2.2	1.5	1.2	1.4	1.2	1.7	1.1	1.5	1.8
Al_2O_3	15.7	16.2	14.5	13.9	16.7	18.1	16.1	16.6	17.5	17.5	15.4	14.6
FeO^t	12.1	10.8	12.4	10.5	9.2	9.6	8.9	10.2	9.4	9.6	11.6	8.7
MnO	0.3	0.2	0.2	0.3	0.5	0.4	0.6	0.4	0.3	0.5	0.3	0.5
MgO	7.2	6.4	4.6	4.8	5.2	4.0	4.6	4.3	3.0	3.5	4.2	3.1
CaO	2.9	9.6	6.7	8.6	4.5	7.3	7.1	7.8	6.3	6.0	5.3	7.4
Na_2O	3.8	3.0	4.0	3.3	4.1	4.6	3.5	3.5	3.8	4.3	3.8	2.1
K_2O	2.1	0.8	0.3	0.7	0.7	0.7	0.7	0.7	1.7	1.1	0.7	1.2
P_2O_5	0.4	0.2	0.2	0.4	0.3	0.2	0.3	0.3	0.4	0.2	0.4	0.5
LOI	4.7	2.2	7.1	1.2	3.1	1.8	1.8	1.5	2.3	1.6	1.5	3.7
TOTAL	99.4	99.5	100.2	96.1	97.2	99.4	97.2	99.1	99.4	98.5	97.9	97.7
Rb	58	28	6	30	50	25	57	22	22	32	10	36
Sr	264	518	443	451	436	388	412	109	371	562	550	569
Y	30	29	34	36	25	22	36	22	33	21	39	30
Zr	151	97	84	70	89	58	232	74	94	69	88	64
Nb	8	7	4	8	5	4	5	4	16	5	5	5
Ba	714	412	141	397	327	401	396	547	721	124	380	953
Ni	99	50	62	35	19	25	30	22	24	43	12	11
Cr	270	211	237	180	83	191	99	109	131	133	92	93
V	359	282	407	370	371	340	326	312	234	233	329	320
Ga	17	17	21	19	19	20	20	15	20	20	18	20
Sc	n.d.	46	56	49	n.d.	n.d.	n.d.	n.d.	22	33	n.d.	n.d.
Hf	n.d.	3	2	5	n.d.	n.d.	n.d.	n.d.	7	3	n.d.	n.d.
La	n.d.	11.3	8.6	18.1	n.d.	n.d.	n.d.	n.d.	36.1	12	n.d.	n.d.
Ce	n.d.	30	24	45	n.d.	n.d.	n.d.	n.d.	74	30	n.d.	n.d.
Nd	n.d.	10	10	20	n.d.	n.d.	n.d.	n.d.	40	10	n.d.	n.d.
Sm	n.d.	4.6	3.4	6.8	n.d.	n.d.	n.d.	n.d.	8.3	4	n.d.	n.d.
Eu	n.d.	1.9	1.2	2.2	n.d.	n.d.	n.d.	n.d.	2.1	1.3	n.d.	n.d.
Yb	n.d.	3	3	4	n.d.	n.d.	n.d.	n.d.	4	2	n.d.	n.d.
Lu	n.d.	0.47	0.48	0.67	n.d.	n.d.	n.d.	n.d.	0.53	0.33	n.d.	n.d.

	Mafic Dykes									
	E16-722	E16-723	E16-727	E16-719	E16-729	E16-544	E16-234	E16-201	E16-203	E16-200
SiO_2	54.6	56.3	57.0	57.3	57.3	57.6	49.9	49.9	51.5	51.6
TiO_2	1.1	0.9	1.1	1.4	0.9	0.7	1.1	0.9	0.9	1.3
Al_2O_3	16.8	15.4	14.5	15.2	15.5	14.1	17.2	16.6	17.9	17.3
FeO^t	9.6	8.3	8.7	8.0	7.8	10.5	9.4	10.6	8.3	9.3
MnO	0.5	0.4	0.6	0.5	0.4	0.2	0.3	0.4	0.4	0.2
MgO	3.5	3.3	2.2	3.0	3.2	1.4	5.2	7.3	5.0	5.3
CaO	6.5	9.2	4.6	4.6	6.8	12.8	6.8	5.6	8.2	7.5
Na_2O	4.2	2.3	2.1	2.9	3.2	3.3	4.5	2.6	3.0	2.9
K_2O	0.8	0.2	2.0	1.1	0.7	0.1	0.7	2.2	0.8	0.9
P_2O_5	0.2	0.2	0.4	0.3	0.2	0.2	0.2	0.2	0.3	0.4
LOI	1.4	3.5	2.2	2.8	3.5	3.9	3.4	3.6	2.7	3.6
TOTAL	99.2	100.0	95.4	97.1	99.5	104.8	98.7	99.9	99.0	100.3
Rb	16	18	50	37	8	25	19	102	47	52
Sr	336	409	442	356	338	440	629	428	560	526
Y	23	26	20	37	19	25	24	14	14	24
Zr	81	71	100	103	71	101	58	48	54	99
Nb	3	6	4	5	4	4	5	4	5	9
Ba	445	61	1675	767	362	29	445	704	320	379
Ni	44	5	< 5	9	52	< 5	11	32	17	27
Cr	222	54	160	43	24	120	193	285	238	113
V	225	153	134	208	167	245	190	244	149	250
Ga	20	18	19	19	17	20	18	17	17	18
Sc	32	29	n.d.	30	29	18	30	n.d.	n.d.	n.d.
Hf	3	3	n.d.	4	3	1	2	n.d.	n.d.	n.d.
La	12.3	13	n.d.	15.9	12	12.5	12.6	n.d.	n.d.	n.d.
Ce	31	31	n.d.	42	29	30	30	n.d.	n.d.	n.d.
Nd	10	10	n.d.	20	20	20	10	n.d.	n.d.	n.d.
Sm	4.1	3.8	n.d.	6	3.8	3.8	3.8	n.d.	n.d.	n.d.
Eu	1.3	1.4	n.d.	2.2	1.3	2.2	1.5	n.d.	n.d.	n.d.
Yb	2	2	n.d.	4	2	2	2	n.d.	n.d.	n.d.
Lu	0.36	0.36	n.d.	0.57	0.37	0.35	0.05	n.d.	n.d.	n.d.

n.d. = not determined

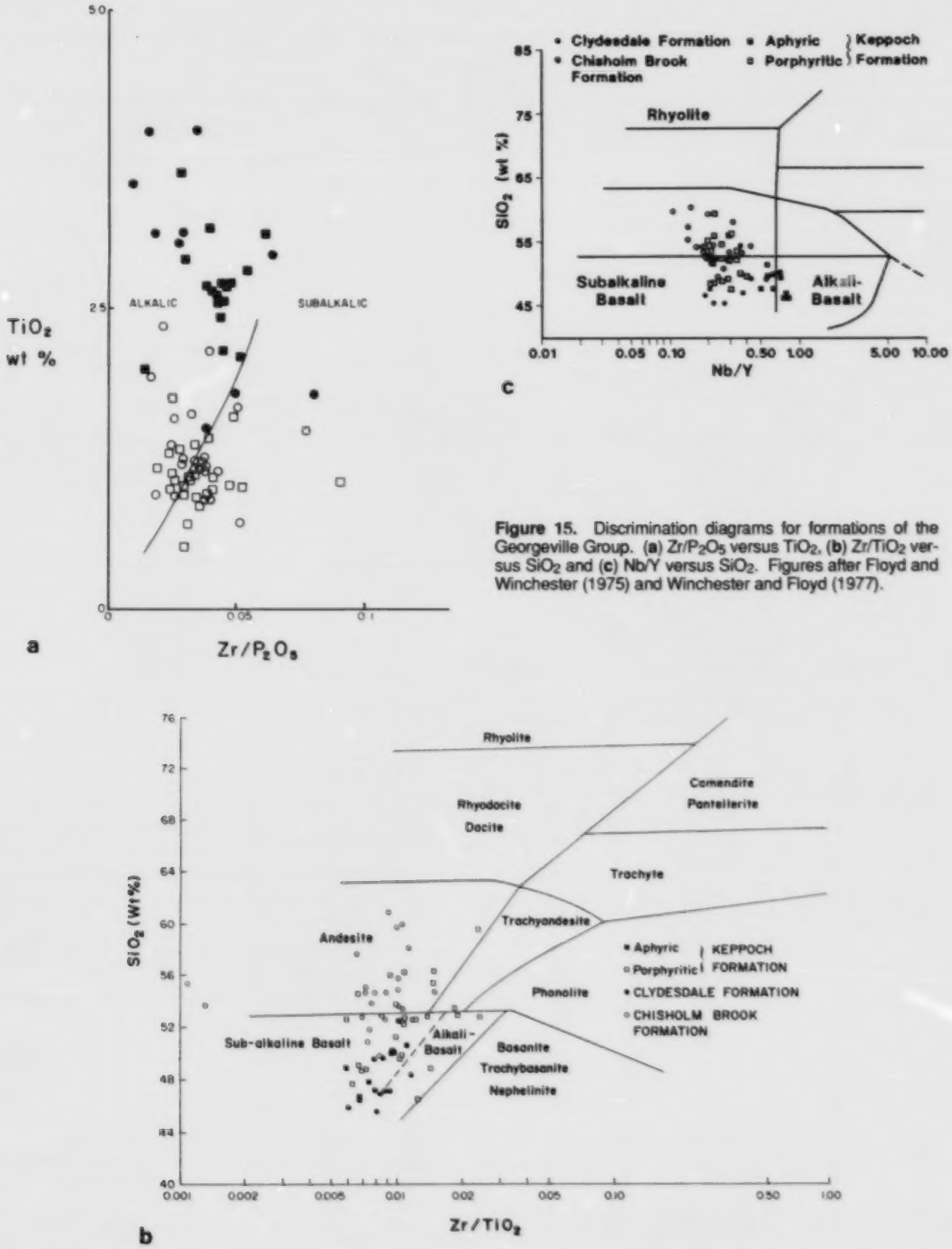


Figure 15. Discrimination diagrams for formations of the Georgeville Group. (**a**) Zr/P_2O_5 versus TiO_2, (**b**) Zr/TiO_2 versus SiO_2 and (**c**) Nb/Y versus SiO_2. Figures after Floyd and Winchester (1975) and Winchester and Floyd (1977).

a negative Nb anomaly, are typical of subduction-related volcanic rocks. On the FeO^t-MgO-Al_2O_3 diagram (Fig. 19), most samples plot in the ocean floor or calc-alkalic fields. On various discrimination diagrams involving Ti, Y, Zr and Cr they predominantly plot in the volcanic arc basalts or ocean floor fields (Fig. 20a-d after Pearce and Cann, 1973; Pearce and Norry, 1979; Meschede, 1986). The concentrations of Ti and V in these rocks overlap the abundance ranges of basalts from marginal basins and calc-alkalic basalts (Pearce and Cann, 1973; Shervais, 1982). The Ti/V ratios of these rocks resemble those of marginal basins.

In summary, the low Ti basalts and andesites have characteristics of both island arc calc-alkalic and ocean floor suites. Weaver et al. (1979) and Saunders and Tarney (1984) attribute similar relationships in modern environments (i.e. tholeiitic-calc-alkalic, ocean floor-orogenic) to contamination of the mantle source region in the early stages of rifting in a back arc environment.

The stratigraphic position of the basalts (i.e. at the base of a sequence which documents the progressive development of a sedimentary basin) suggests that the Chisholm Brook Formation may represent incipient stages of rifting within an island arc.

Geochemistry of the Clydesdale Formation

Magmatic affinity. Basalts of the Clydesdale Formation may be distinguished from the basalts and andesites of the Chisholm Brook Formation by the generally lower Si and Al contents (Fig. 21) and the higher concentrations of Fe (Fig. 16b), P, Ti, Zr (Fig. 20a) and REE (Fig. 17b). The chemistry of the basalts shows an affinity with both alkalic and tholeiitic suites. The humped multielement pattern of Figure 18, which displays an enrichment in elements from Ce to Ti, is typical of alkalic or continental tholeiitic basalts (Pearce, 1982). The high Fe and positive correlations between Fe, Ti, P and Zr are typical of tholeiitic ferrobasalts. An alkalic affinity is

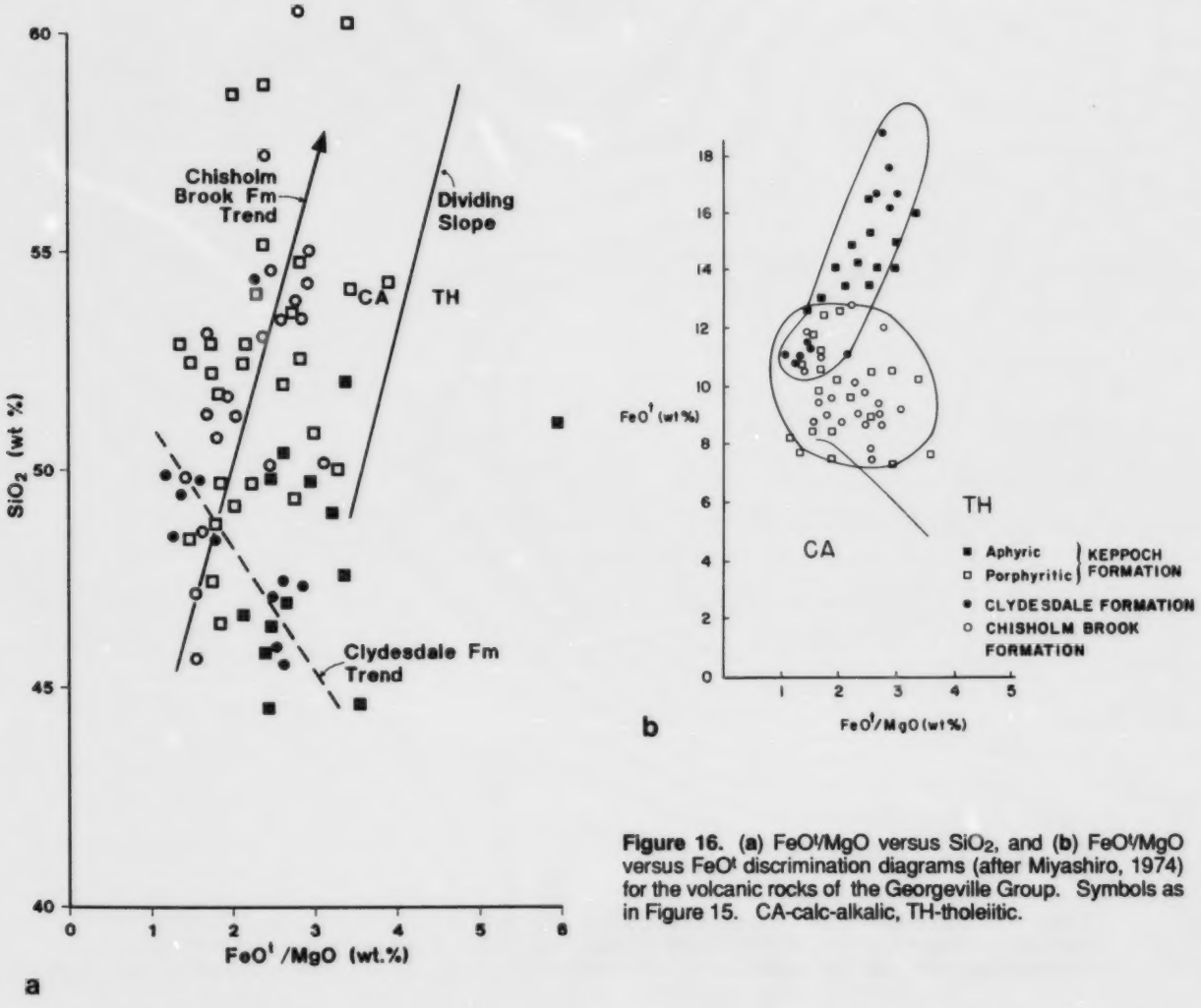


Figure 16. (**a**) FeO^t/MgO versus SiO_2, and (**b**) FeO^t/MgO versus FeO^t discrimination diagrams (after Miyashiro, 1974) for the volcanic rocks of the Georgeville Group. Symbols as in Figure 15. CA-calc-alkalic, TH-tholeiitic.

suggested by the Zr/P_2O_5 versus TiO_2 and SiO_2 versus Zr/TiO_2 diagrams (Fig. 15a,b). However, continental tholeiites that have undergone extensive fractionation involving Fe and Ti enrichment tend to plot in the alkalic fields. The moderate to strong LREE enrichment (La/Sm from 2 to 8, and La/Yb from 5 to 23) fall into the range of continental tholeiites or alkalic suites (Fig. 17b). Some samples have a very high Nb/Y ratio (i.e. >2), which is typical of alkalic basalts; others have a low Nb/Y ratio and plot in the subalkaline field (Fig. 15c). These rocks have distinctly fractionated REE patterns (La/Yb <20) and MORB-normalized multielement patterns that are typical of alkalic suites.

Tectonic setting. The basalts plot in the within-plate field on the TiO_2 versus Zr diagram (Fig. 20a), and the high Zr/Y ratio (between 4.5 and 10) is also consistent with a within-plate tectonic setting (Fig. 20b). The within-plate tectonic setting and the generally tholeiitic-alkalic character of the Clydesdale Formation, together with its lithological association (i.e. these rocks are interlayered with turbidites) and the wide ranges in the abundances of TiO_2, FeO^t and high Zr (Table 2), are typical of an intra-continental rift setting. Multielement profiles lie between supra-subduction zone and nonsupra-subduction zone basalts (Fig. 18). According to Pearce et al. (1984b) the anomalously low Nb concentration relative to most mafic suites is typical of basalts genetically related to subduction zones. Thus, the interpretation of the tectonic

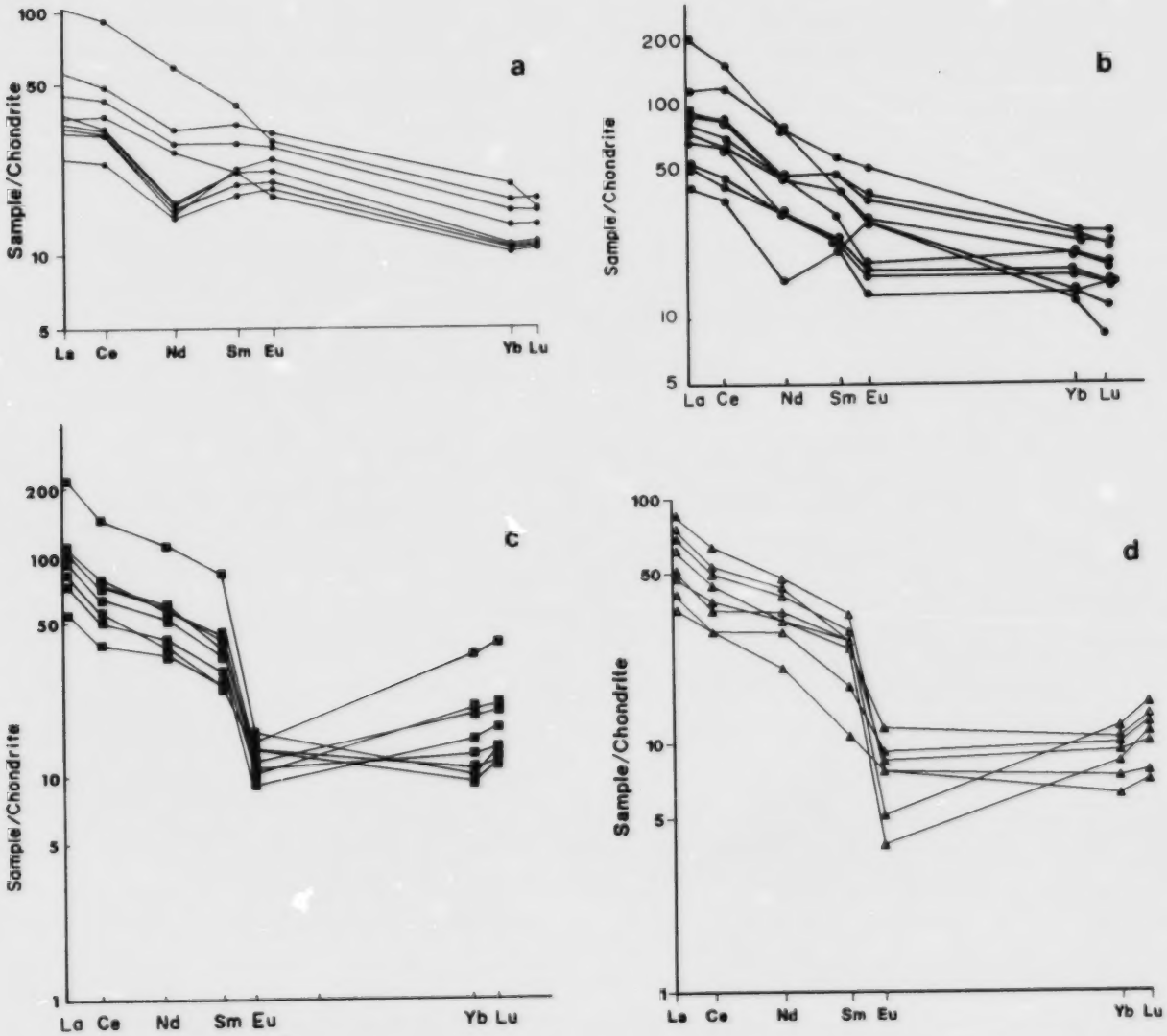


Figure 17. REE diagrams for the (**a**) Chisholm Brook, (**b**) Clydesdale and (**c**) to (**e**) Keppoch formations (basalts, basaltic andesites and felsic volcanic rocks, respectively). (**f**) shows the field of basalts derived from enriched mantle (P-type, vertical lines) and ocean islands (stipples) (see Le Roex et al., 1983).

environment of the Clydesdale Formation is compatible with that of the underlying Chisholm Brook Formation and suggests that the basin in which the Clydesdale Formation was deposited in an intra-continental marine basin genetically associated with subduction.

Geochemistry of the Keppoch Formation

The relatively low SiO_2 and high MgO and FeO^t confirm the mafic character of the basalts and the mafic to intermediate character of the basaltic andesites of this formation. The SiO_2 concentration of the felsic rocks indicates that they vary from rhyodacite to rhyolite in composition. Mafic rocks are highly variable in chemistry. For example, for similar values of SiO_2, TiO_2 varies from 1.0 to 4.5 weight percent, Zr from 40 to 380 ppm and Cr from 5 to 190 ppm. The basalts may be distinguished from the basaltic andesites by the generally lower Si and Al contents (Fig. 21) and the higher Fe (Fig. 16b), P, Ti, Zr (Fig. 20a) and LREE values (Fig. 17c,d). The generally lower Ni as well as lower Ti, Zr and REE values in the basaltic andesites suggest that the basaltic andesites and basalts are not comagmatic. Further, on standard discrimination diagrams (Fig. 15, 16 and 20) they display different magmatic and tectonic affinities. Therefore, chemically the rocks of this formation can be divided into two groups, which correspond to porphyritic and aphyric types distinguished on the basis of petrography. The geochemistries of these two rock types are discussed separately below.

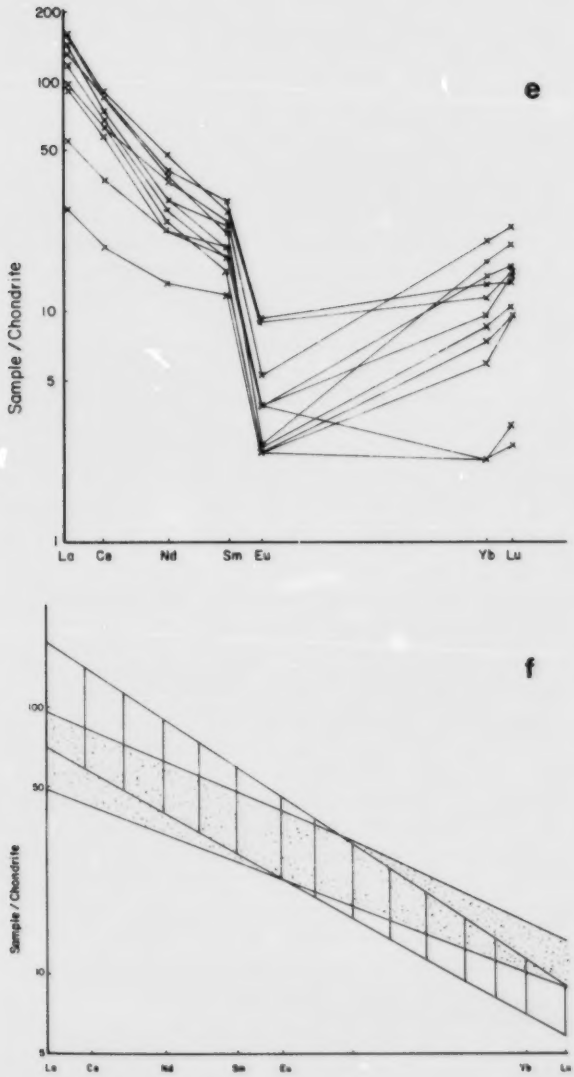


Figure 17. (cont.)

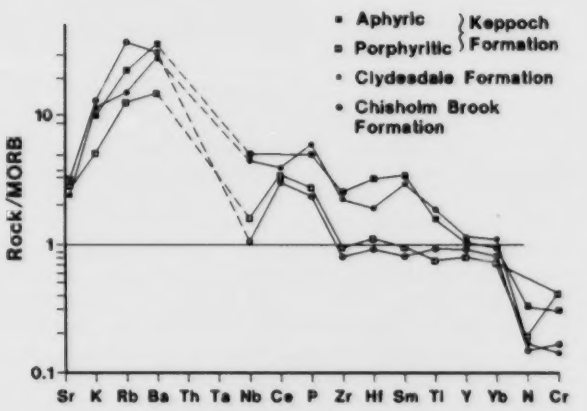


Figure 18. Multielement plot with the average abundances for mafic volcanic rocks normalized to MORB for the Chisholm Brook, Clydesdale and Keppoch formations. MORB values from Pearce et al., 1984b. Symbols as in Figure 15.

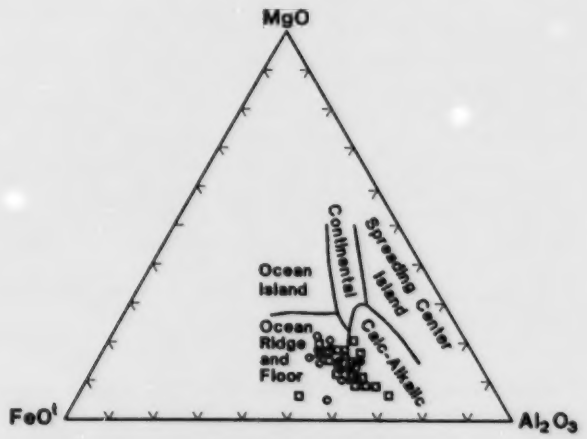


Figure 19. FeO^t-MgO-Al_2O_3 diagram (after Pearce et al., 1977) for the basaltic andesites of the Chisholm Brook and Keppoch formations. Symbols as in Figure 15.

Basalts (aphyric flows)

Magmatic affinity. The geochemistry of the basalts displays characteristics of both tholeiitic and alkalic suites and is summarized in Figures 15 and 16. The basalts consist of 46 to 53 weight percent SiO_2. They display distinct variation trends, characterized by a strong enrichment of Fe and Ti with a simultaneous slight depletion in Mg and possibly Ca in differentiated rocks. The concentration of FeO^t in some samples reaches greater than 15 percent, and TiO_2 is greater than 3.5 percent. The major element variations are typical of continental tholeiites that underwent fractionation of a gabbroic assemblage of clinopyroxene + plagioclase ± olivine (Cox, 1980). The humped multielement pattern and enrichment in elements from Ce to Ti is typical of continental tholeiitic and alkalic suites. On discrimination diagrams, the basalts plot in the alkalic field (Fig. 15a,b). This, however, could be due to Ti enrichment during differentiation of a tholeiitic magma. A continental tholeiite or alkalic chemistry is also consistent with the moderate to high La/Yb ratio and the LREE enrichment (Fig. 17c).

Tectonic setting. On the basis of TiO_2 and Zr abundances (Fig. 20a), the high Zr/Y ratio (between 4 and 11, Fig. 20c), and other conventionally used diagrams such as Ti/100 versus Zr versus 3Y (Fig. 20b), Ti/Y versus Nb/Y and Nb versus Zr versus Y, the basaltic rocks are interpreted to have a within-plate tectonic setting. The lithological association (i.e. these rocks were deposited in a subaerial environment) indicates a continental setting. The multielement profile lies between supra-subduction zone and nonsupra-subduction zone basalts (Fig. 18). However, the anomalously low Nb is consistent with basalts that have a genetic association with subduction (Pearce et al., 1984b). The wide ranges in Zr/Y, Zr/Nb and Zr/La (Fig. 22) suggest that the parental magmas were derived from a heterogeneous upper mantle source.

Basaltic andesites (porphyritic flows)

Magmatic affinity. Basaltic andesites display a subalkalic chemistry (Fig. 15b). Although these rocks have a relatively high iron content (FeO^t averaging between 10 and 11 weight percent), the suite does not show an iron enrichment trend on the FeO^t/MgO versus FeO^t and FeO^t/MgO versus SiO_2 diagrams (Fig. 16a,b; after Miyashiro, 1974). This indicates that these rocks have calc-alkalic tendencies (Fig. 16a). The decrease of Ti and V with Si, the low abundances of transition elements (particularly Ni and Cr), the abundances of Sr, Ba and La and some element ratios such as Zr/Nb, Ti/Y, La/Nb and Nb/Y are comparable to calc-alkalic suites. The REE pattern (Fig. 17d) displays a moderate LREE enrichment and falls within the range of volcanic arc calc-alkalic basalts (Gill, 1981).

Tectonic setting. On the FeO^t-MgO-Al_2O_3 diagram, the samples plot in the ocean floor and calc-alkalic fields (Fig. 19). Ti, Zr, Cr, Y and Zr/Y abundances are consistent with a volcanic arc or ocean floor setting (Fig. 20a,c,d) and are inconsistent with a within-plate environment. Many geochemical features are atypical of ocean floor rocks but are typical of a volcanic arc setting such as the extent of differentiation as indicated by the range in high SiO_2 (Fig. 15b), the limited range in FeO^t/MgO (Fig. 16a), the enriched multielement pattern relative to MORB (Fig. 18), the low Ni and Cr contents (Table 3) and the relatively high REE and moderate LREE enrichment (Fig. 17d). These characteristics are very similar to those found for the Chisholm Brook Formation. An oceanic environment is inconsistent with the lithological association, which indicates that these basalts were deposited in a subaerial continental environment. In summary, the geochemistry of the basaltic andesites shows many of the characteristics of calc-alkalic suites, although some evidence of transitional to ocean floor tholeiites is also indicated. The similarity of the geochemistry and stratigraphic position of these basaltic andesites with those of the Chisholm Brook Formation suggests a similar origin, i.e. they may represent volcanism in the incipient stages in the development of a intra-arc basin.

Felsic rocks

In comparison to the mafic rocks, the felsic volcanic rocks are higher in SiO_2, K_2O and Na_2O, and are lower in TiO_2, MgO, CaO, Ni and Cr (Table 3c). They are metaluminous to peraluminous (all iron calculated as ferrous) and range from rhyodacite to rhyolite in composition. Normalized multielement variation diagrams for these rocks are shown on Figure 23 (after Pearce et al., 1984a). The plot shows enrichment in K, Rb, Ba and Th and depletion in the incompatible high ionic potential elements relative to the ocean ridge granites. According to Pearce et al. (1984a), these trends are typical of volcanic arc rocks of granitic composition. The concentrations of Ce, Zr and Hf are generally too high to be compatible with a collisional or postcollisional setting, whereas the concentrations of Nb, Ce, Hf, Zr, Sm, Y and Yb are too low for a within-plate setting (see Fig. 12b,c of Pearce et al., 1984a).

Comparison between the volcanic rocks of the northern and southern Antigonish Highlands

The basaltic andesites of the Keppoch Formation have a similar geochemical signature to those of the Chisholm Brook Formation of the northern highlands. Both suites are subalkalic and have many similarities to modern calc-alkalic suites, although some transition to ocean floor basalts is indicated. Stratigraphic considerations (Fig. 3 and 11) and geochronological data indicate that the basaltic andesites of the Chisholm Brook and Keppoch formations are roughly coeval. Their similar and distinctive chemistry suggests that they are probably comagmatic and that they represent the incipient stages in the development of a back-arc basin.

The basalts of the Keppoch Formation compare favourably with the Clydesdale Formation and the high Ti rocks of the Chisholm Brook Formation. These suites have tholeiitic and alkalic tendencies and a within-plate tectonic setting, and display remarkably similar chemistries. The basalts of the Chisholm Brook and Keppoch formations are separated from the Clydesdale Formation stratigraphically by at least 1700 m

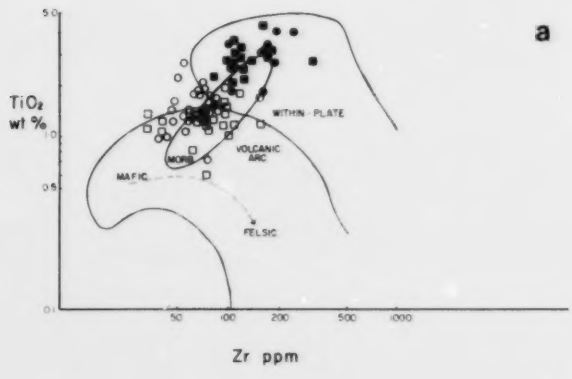
of turbidites. A study of the vertical gradient magnetic anomalies in the Antigonish Highlands (Fig. 10) shows that the Clydesdale Formation forms part of an elongate anomaly that extends approximately north-south and is oblique to contacts between the Keppoch Formation and the overlying sediments. These relationships suggest that the basalts may be comagmatic and that they represent a volcanic ridge. The different stratigraphic levels at which this activity is found may be attributed to volcanic activity concentrated along a major line of weakness through which mafic magmas preferentially erupted.

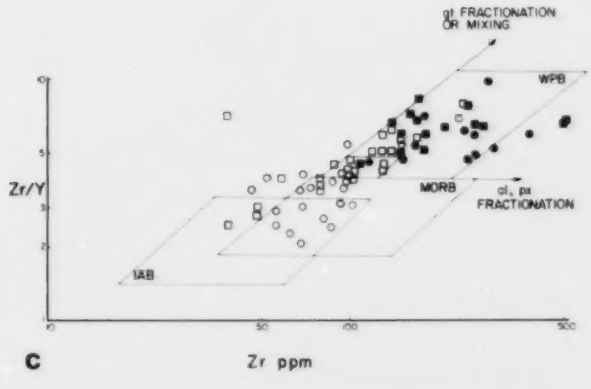
Petrogenesis

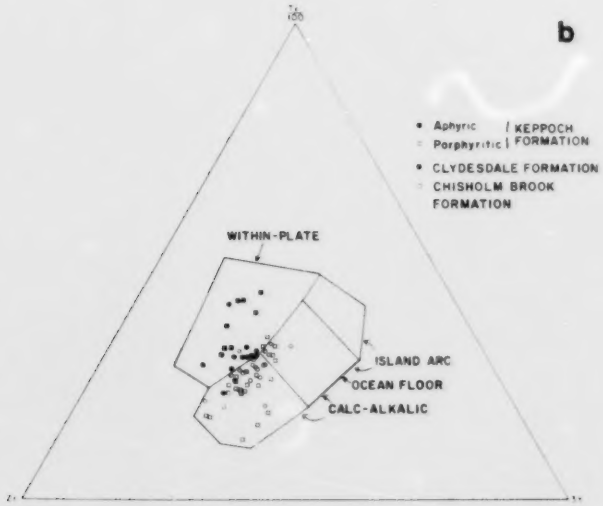
Basaltic andesites

The remarkable similarity between the basaltic andesites of the Chisholm Brook and Keppoch formations suggests that they had a similar petrogenesis. Thus their petrology is discussed together.

The low Ni contents of these rocks indicate that the basaltic andesites are fractionated with respect to olivine; the range in Cr together with the presence of augite







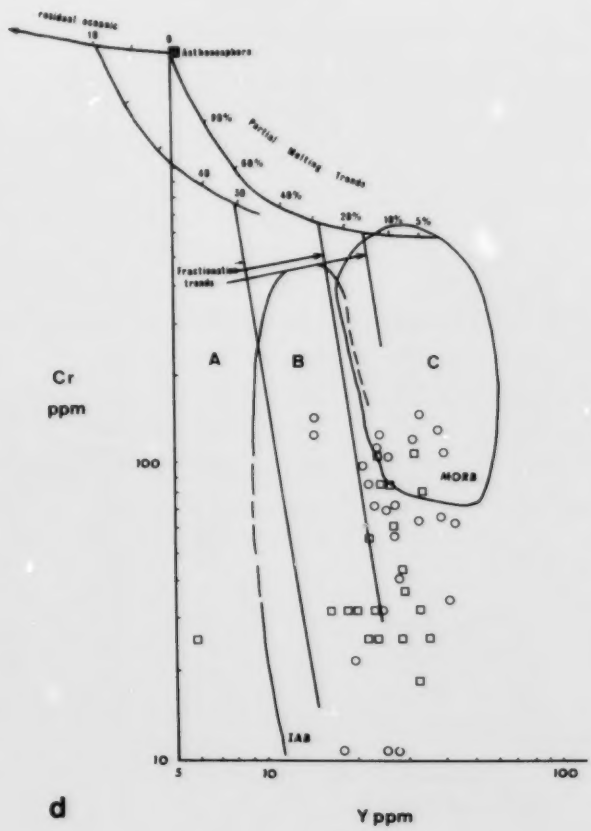


Figure 20. (**a**) TiO_2 versus Zr, (**b**) Ti/100 versus Zr versus 3Y, (**c**) Zr/Y versus Zr for all volcanic rocks and (**d**) Cr versus Y for the mafic volcanic rocks of the Chisholm Brook, Clydesdale and Keppoch formations. Symbols as in Figure 15. Diagrams after Pearce et al. (1981), Pearce and Cann (1973) and Pearce and Norry (1979). WPB-within-plate basalts, MORB-mid-ocean ridge basalts, IAB-island arc basalts, gt-garnet, ol-olivine, px-pyroxene.

microphenocrysts suggests that augite was on the liquidus. Olivine and augite fractionation is also consistent with Fe and moderate LREE enrichment. Although plagioclase phenocrysts also occur, geochemical trends related to plagioclase fractionation are difficult to decipher due to the mobility of Ca and Sr during alteration. The absence of a Eu anomaly in the Chisholm Brook rocks and the presence of an anomaly in the Keppoch rocks indicates that plagioclase fractionation was important in the Keppoch Formation only. Cr-Y relationships (Fig. 20d) indicate that the basaltic andesites may represent between 5 and 40 percent partial melting. However, the Keppoch Formation, with its generally lower Cr contents, is probably more extensively fractionated.

The calc-alkalic affinity suggests the proximity of a subduction zone. The large ion lithophile element (LIL) enrichment may be attributed to progressive contamination of the mantle wedge source area either by calc-alkalic magma or by LIL-enriched fluids derived from dehydration of the subduction zone (e.g. Saunders and Tarney, 1984). It is difficult to distinguish between these possibilities because both yield magmas with calc-alkalic tendencies. However, both models are consistent with a genetic relationship with subduction. In each case, during the progressive melting of the mantle

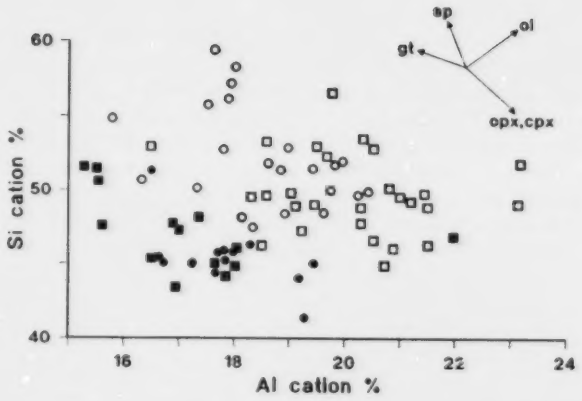


Figure 21. Si versus Al cation plot for the mafic volcanic rocks of the Georgeville Group. Symbols as in Figure 15.

Table 2. Geochemistry of the Clydesdale Formation

	10157	10104	10156	10100	10145	10147	10120	10162	10163	10154
SiO_2	45.60	45.80	46.90	47.10	47.10	48.40	49.70	49.70	49.80	54.70
TiO_2	3.90	3.07	3.00	3.10	2.90	2.60	1.50	1.80	3.70	1.80
Al_2O_3	15.50	15.40	14.70	15.30	14.60	17.80	16.60	16.70	18.50	14.90
Fe_2O_3	18.60	17.70	16.80	16.00	16.40	10.70	10.80	11.10	11.16	11.20
FeO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	6.90	6.30	6.10	5.80	5.60	7.50	8.60	7.40	6.90	5.40
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CaO	3.90	5.50	7.20	7.20	8.20	6.20	7.90	7.90	8.50	6.10
Na_2O	2.80	2.90	2.80	3.20	2.90	3.40	3.80	3.80	4.50	3.40
K_2O	1.40	1.80	1.40	1.50	1.40	2.70	0.90	n.d.	0.70	2.30
P_2O_5	0.90	n.d.	0.90	0.90	0.40	0.70	0.30	0.30	1.90	0.30
H_2O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CO_2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
TOTAL	99.50	99.47	99.80	100.10	99.50	100.00	100.10	99.70	105.66	100.10
Rb	85	28	35	38	16	60	15	16	13	65
Sr	198	322	372	343	271	293	358	493	251	273
Y	59	26	54	55	45	30	26	33	33	40
Zr	306	181	249	264	263	293	118	153	153	245
Nb	14	12	11	11	16	88	11	18	18	14
Ba	1067	510	925	1065	329	888	272	306	215	985
Ni	51	21	53	48	69	105	103	90	72	46
Cr	36	74	29	38	100	387	206	229	212	89
V	444	102	335	346	337	329	264	304	350	244
Ga	23	23	22	24	24	21	16	21	23	23
Sc	49	16	47	48	45	35	36	43	43	30
Hf	6	5	5	6	5	6	n.d.	4	3	5
La	40	25	27	31	23	70	14	18	19	27
Ce	107	56	61	74	57	132	31	40	38	61
Nd	50	20	30	30	30	50	10	20	20	30
Sm	11.9	4.7	6.5	3.0	2.2	8.6	4.4	5.4	5.5	6.5
Eu	4.0	n.d.	2.8	10.3	8.1	2.2	2.2	1.3	1.2	1.4
Yb	5.9	2.9	4.4	5.6	4.4	3.0	2.7	3.6	3.7	4.4
Lu	0.9	0.5	0.6	0.8	0.6	0.4	0.3	0.5	0.5	0.6

n.d. = not determined

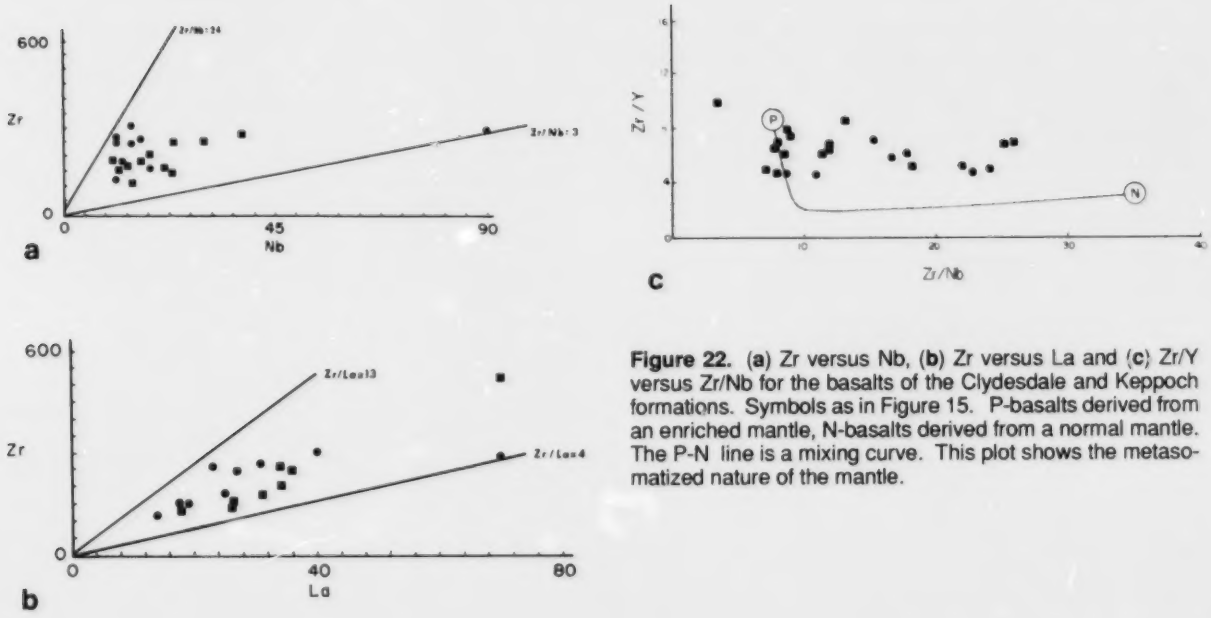


Figure 22. (**a**) Zr versus Nb, (**b**) Zr versus La and (**c**) Zr/Y versus Zr/Nb for the basalts of the Clydesdale and Keppoch formations. Symbols as in Figure 15. P-basalts derived from an enriched mantle, N-basalts derived from a normal mantle. The P-N line is a mixing curve. This plot shows the metasomatized nature of the mantle.

Table 3a. Geochemistry of the aphyric mafic rocks (basalts) of the Keppoch Formation

	56-5C	49-3C	64-1A	56-3B	64-1B	42-4F	55-2C1	26-1F1	24-22A
SiO_2	48.9	44.3	48.3	44.6	48.3	45.8	48.5	46.3	47.8
TiO_2	2.35	2.75	2.97	4.13	2.58	2.42	1.93	2.03	1.93
Al_2O_3	14.8	15.2	13.4	15.2	14.5	15.1	19.3	15.4	14.6
Fe_2O_3!	13.6	13.7	14.6	16.3	13.8	14.1	10.0	12.6	13.4
MgO	5.08	6.13	4.68	6.42	4.70	6.48	2.96	7.70	5.62
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CaO	7.41	8.87	6.23	3.68	7.22	7.26	6.91	6.53	5.91
Na_2O	3.29	2.83	4.06	2.65	3.23	2.38	3.98	2.37	2.41
K_2O	1.99	0.98	1.62	1.99	1.61	1.63	2.12	1.51	2.34
P_2O_5	0.40	0.53	0.43	0.50	0.46	0.33	1.06	0.23	0.32
LOI	n.d.	3.94	1.93	4.23	2.70	2.62	2.47	3.70	4.08
TOTAL	97.82	99.23	98.22	99.70	99.10	98.12	99.23	98.37	98.41
Rb	48	16	20	60	48	56	53	953	99
Sr	403	53	140	212	355	527	606	317	270
Y	31	26	43	37	36	26	33	24	36
Zr	207	151	280	234	221	176	173	116	209
Nb	22	20	37	31	20	21	21	16	25
Ba	697	400	510	609	577	590	948	992	338
Ni	40	59	18	26	36	51	6	109	44
Cr	50	99	20	10	29	42	6	263	119
V	332	279	370	401	344	303	159	328	306
Ga	22	19	1	24	20	22	22	19	23
Sc	34	33	39	34	33	32	n.d.	n.d.	n.d.
Hf	5	3	9	7	7	6	n.d.	n.d.	n.d.
La	31	233	37	36	34	26	n.d.	n.d.	n.d.
Ce	54	43	64	63	61	48	n.d.	n.d.	n.d.
Nd	32	26	34	35	37	24	n.d.	n.d.	n.d.
Sm	7.1	6	9.1	8.3	7.8	6.3	n.d.	n.d.	n.d.
Eu	0.7	1	0.8	0.8	0.8	1	n.d.	n.d.	n.d.
Yb	3.1	2.1	4.1	2.8	3.3	2.4	n.d.	n.d.	n.d.
Lu	0.6	0.4	0.7	0.4	0.5	0.4	n.d.	n.d.	n.d.

n.d. = not determined

wedge, the basaltic andesites became progressively enriched in LIL elements producing a typical calc-alkalic trace element signature. The low rate of increase of FeO^t/MgO in these rocks may be attributed to the oxidation of the source area (e.g. Miyashiro, 1974) resulting in fractionation of magnetite. Wide ranges in petrogenetic-indicating ratios such as Zr/Nb and Zr/Y (e.g. Fig. 20c) are most satisfactorily explained by contamination of the mantle source (Pearce and Norry, 1979).

Basalts

The petrology of the basalts of the Keppoch and Clydesdale formations is discussed together because of the general similarity in chemistry. In the Keppoch Formation, Al shows a negative correlation with Si and a positive correlation with Mg (Fig. 21 and 24). These variations could result from spinel ± garnet ± plagioclase fractionation, whereas the Clydesdale rocks appear to be dominated by garnet, olivine and

Table 3b. Geochemistry of the basaltic andesites of the Keppoch Formation. CO_2 values (in brackets) are included in LOI

	589	641	19A-11	58-5A	17-9	54-4B1	54-4B12	19-6A	58-5D	57-6E	58-3B	61A-3A	61-4A	63-1A	43-11	17-8
SiO_2	51 4	44 9	50 4	51 6	46 4	49 6	54 4	46 9	47 9	47 3	46 8	47 3	54 8	46 2	51 9	54 9
TiO_2	1 07	1 24	1 05	0 89	1 35	1 09	1 07	0 94	1 15	1 84	1 83	1 55	1 01	1 30	0 67	1 01
Al_2O_3	19 0	18 2	16 4	16 5	18 1	18 8	18 0	18 1	18 1	16 0	16 6	16 2	17 3	18 3	19 9	16 4
Fe_2O_3	9 38	12 1	7 90	8 14	13 1	9 92	9 71	10 8	11 0	12 4	10 8	11 5	9 60	10 9	7 67	8 72
MgO	3 72	6 89	5 05	6 93	7 68	3 36	4 53	6 50	5 39	6 73	9 14	6 99	3 31	6 82	1 97	4 56
MnO	0 12	0 23	0 39	0 12	0 34	0 23	0 25	0 22	0 19	0 37	0 21	0 29	0 16	0 20	0 15	0 22
CaO	4 07	6 16	5 04	7 30	1 84	7 78	4 02	8 95	7 01	7 58	3 66	8 19	4 55	9 01	3 91	3 80
Na_2O	4 81	3 43	3 87	2 71	2 40	2 89	3 70	2 15	3 92	2 50	3 45	2 61	4 70	2 56	3 88	3 83
K_2O	3 20	0 79	3 13	2 40	3 72	2 66	0 90	1 15	1 54	2 02	1 45	1 74	1 57	1 27	3 34	3 07
P_2O_5	0 39	0 29	0 34	0 43	0 27	0 26	0 23	0 19	0 30	0 26	0 30	0 38	0 25	0 36	0 28	0 32
BaO	0 16	0 07	0 31	0 10	0 04	0 08	0 06	0 04	0 06	0 05	0 09	0 07	0 15	0 07	0 04	0 07
LOI	2 08	4 70	5 77	2 77	4 70	2 47	3 00	3 54	2 85	2 93	4 77	2 85	1 77	2 82	5 62	3 08
CO_2	(0 1)	(0 6)	(3 0)	(0 2)	(0 4)	(0 1)	(0 1)	(0 1)	(0 3)	(0 3)	(0 2)	(0 1)	(0 1)	(0 1)	(2 8)	(0 7)
TOTAL	99 5	99 1	99 7	100 0	100 0	99 2	100 0	99 6	99 5	100 1	99 2	99 7	99 3	99 9	99 4	99 8
Rb	100	30	90	60	230	100	50	40	70	50	50	60	40	30	100	140
Sr	370	520	220	630	70	550	520	660	660	370	330	410	540	770	140	330
Y	25	16	21	22	22	22	21	18	22	27	33	31	27	23	24	27
Zr	110	60	80	140	100	80	90	50	100	130	150	170	130	130	80	150
Nb	1	4	14	6	1	5	3	1	7	7	3	9	5	8	4	4
Ba	1588	624	3043	n d	373	793	513	401	584	n d	n d	813	1579	n d	372	n d
Ni	15	16	32	80	12	12	14	28	16	44	60	88	16	62	6	22
Cr	10	40	80	190	40	30	30	640	30	80	160	130	30	140	10	50
V	130	397	260	n d	470	282	290	310	296	n d	240	224	300	n d	88	n d
Ga	22	21	15	n d	23	21	18	18	19	n d	n d	21	7	n d	15	n d
Sc	25 0	39 6	35 5	n d	n d	35 9	n d	40 7	n d	n d	35 4	n d	31 9	n d	n d	n d
La	22 5	13 0	25 4	n d	n d	21 3	n d	12 0	n d	n d	22 0	n d	24 0	n d	n d	n d
Ce	40 0	30 0	47 0	n d	n d	39 0	n d	25 0	n d	n d	45 0	n d	42 0	n d	n d	n d
Nd	24 0	18 0	27 0	n d	n d	20 0	n d	18 0	n d	n d	29 0	n d	23	n d	n d	n d
Sm	6 1	4 7	5 4	n d	n d	4 9	n d	3 5	n d	n d	6 4	n d	4 8	n d	n d	n d
Eu	0 6	0 7	0 7	n d	n d	0 3	n d	0 6	n d	n d	0 6	n d	0 9	n d	n d	n d
Yb	2 2	1 9	2 2	n d	n d	1 9	n d	1 6	n d	n d	3 2	n d	2 8	n d	n d	n d
Lu	0 35	0 29	0 43	n d	n d	0 38	n d	0 27	n d	n d	0 5	n d	0 43	n d	n d	n d
Hf	4 0	2 0	3 0	n d	n d	3 0	n d	2 0	n d	n d	5 0	n d	3 0	n d	n d	n d

n d = not determined

Table 3b. (cont.)

	58-2A	58-2E	58-3A5	64-1E	483	553	590	55-5B	26-1G1	567	552	587	501	47-6B	55-4A1	55-2C1
SiO_2	51 1	51 2	51 2	53 4	50 1	51 0	48 1	57 1	47 6	49 5	50 4	51 6	52 3	50 9	53 2	48 5
TiO_2	1 03	0 87	0 98	1 71	1 03	1 18	1 30	0 51	0 94	1 30	1 18	1 02	1 12	2 57	1 90	1 93
Al_2O_3	18 8	16 2	17 4	14 2	17 8	16 7	16 8	17 1	19 0	17 6	16 9	19 1	16 9	12 9	15 2	19 3
Fe_2O_3	9 57	8 04	8 22	12 8	10 2	10 4	10 2	7 36	10 4	11 5	10 2	9 19	10 7	14 3	11 1	10 0
MgO	4 99	6 83	5 84	1 99	5 52	5 92	5 05	2 48	5 15	6 12	6 37	3 33	3 89	2 30	4 77	2 96
MnO	0 21	0 14	0 13	0 62	0 17	0 19	0 18	0 24	0 33	0 17	0 17	0 20	0 15	1 09	0 44	0 33
CaO	3 30	4 48	6 52	5 11	4 03	6 15	10 0	4 27	7 06	2 44	5 21	8 51	6 02	5 33	3 90	8 91
Na_2O	3 48	3 74	4 24	2 46	4 87	2 81	2 89	3 03	3 13	1 34	3 66	3 28	2 93	3 15	4 57	3 96
K_2O	3 97	3 49	1 66	4 95	1 10	2 06	1 47	3 22	2 91	3 74	2 02	1 64	1 31	3 15	1 37	2 12
P_2O_5	0 26	0 04	0 26	0 59	0 21	0 38	0 48	0 37	0 19	0 40	0 37	0 37	0 25	1 08	0 43	1 06
BaO	0 23	0 17	0 07	0 24	0 08	0 06	0 04	0 06	0 14	0 05	0 07	0 07	0 06	0 24	0 06	0 09
LOI	3 39	3 47	3 39	1 70	4 23	2 23	2 47	4 62	3 23	6 08	2 85	1 62	3 38	1 77	2 85	2 47
CO_2	(0 1)	(0 1)	(0 1)	(0 1)	(0 7)	(0 1)	(0 3)	(2 3)	(0 1)	(1 5)	(0 2)	(0 2)	(0 7)	(0 2)	(0 1)	(0 1)
TOTAL	100 2	99 2	100 2	99 8	99 5	99 2	99 1	100 4	100 2	100 3	99 5	100 0	99 1	99 0	99 9	99 7
Rb	140	100	50	80	30	100	30	140	100	160	80	60	40	70	40	50
Sr	560	670	360	200	400	440	790	150	560	50	480	770	640	270	150	630
Y	32	29	24	32	19	30	25	25	8	27	27	25	17	78	41	25
Zr	240	140	100	130	50	130	120	100	40	130	130	100	40	540	150	150
Nb	7	7	3	13	< 1	6	5	< 2	< 1	1	1	5	1	21	25	18
Ba	2283	1716	721	2528	758	579	411	491	1486	410	671	684	584	2580	n d	946
Ni	16	80	62	5	12	22	44	< 2	10	24	28	40	10	< 2	28 0	2
Cr	30	190	140	20	40	40	90	10	30	60	60	10	10	5	40 0	5
V	195	213	218	60	369	302	256	68	318	347	319	219	357	92	180 0	159
Ga	16	16	15	21	18	20	16	13	19	18	19	20	20	25	n d	22
Sc	n d	n d	n d	n d	n d	33 0	30 8	n d	33 1	36 3	36 4	24 3	n d	n d	31 8	n d
La	n d	n d	n d	n d	n d	28 3	29 1	n d	13 9	16 2	24 0	21 5	n d	n d	36 6	n d
Ce	n d	n d	n d	n d	n d	55 0	52 0	n d	25 0	33 0	45 0	42 0	n d	n d	66 0	n d
Nd	n d	n d	n d	n d	n d	30 0	29 0	n d	13 0	20 0	29 0	24 0	n d	n d	47	n d
Sm	n d	n d	n d	n d	n d	6 8	6 0	n d	3 4	5 4	8 4	5 6	n d	n d	8 9	n d
Eu	n d	n d	n d	n d	n d	0 4	0 7	n d	0 6	0 9	0 6	0 4	n d	n d	0 9	n d
Yb	n d	n d	n d	n d	n d	2 6	2 6	n d	1 4	2 3	2 6	2 2	n d	n d	3 9	n d
Lu	n d	n d	n d	n d	n d	0 51	0 24	n d	0 25	0 44	0 49	0 35	n d	n d	0 69	n d
Hf	n d	n d	n d	n d	n d	4 0	4 0	n d	2 0	5 0	4 0	3 0	n d	n d	7 0	n d

n d = not determined

plagioclase fractionation. Thus, in a general way, the major element concentrations can be accounted for by crystal-magma fractionation processes.

However, if crystal fractionation is the dominant process in determining trace element concentrations, then incompatible elements such as Zr, Nb and La should exhibit a strong positive correlation and relatively constant ratios. The wide ranges in Zr/Nb and Zr/La exhibited by both the Keppoch and Clydesdale suites (Fig. 22a,b) cannot be attributed to crystal-magma equilibria processes. Pearce and Norry (1979) attribute ranges in Zr/Nb to source heterogeneity. Furthermore, both suites show a wide range in Zr/Y (Fig. 20c). This range is generally attributed to either garnet fractionation or source heterogeneity (e.g. Pearce and Norry, 1979). However, the lack of correlation of Zr/Y with Cr suggests that the range in Zr/Y cannot be explained by garnet fractionation and is probably due to source heterogeneity.

In summary, although major element variations may result from crystal fractionation, crystal-liquid fractionation alone cannot account for the trace element variations because the incompatible element abundances cannot be predicted from the major element concentrations. Similar features in modern basaltic suites are commonly described (e.g. Frey, 1984). The unusual enrichment in incompatible elements is commonly attributed to the introduction of these elements by mantle metasomatism during or prior to magmatism because dry peridotites lack sufficient quantities of these elements to produce high concentrations of these elements with any acceptable degree of partial melting (e.g. Lloyd and Bailey, 1975; Menzies and Murthy, 1980). This is emphasized on a Zr/Y versus Zr/Nb plot (Fig. 22c) in which the range of Zr/Y and Zr/Nb indicate melting of a heterogeneous source and in which many of the analyses lie close to those derived from an enriched (metasomatized) mantle source (Le Roex et al., 1983; Fodor and McKee, 1986).

Table 3c. Geochemistry of the felsic volcanic rocks of the Keppoch Formation. CO_2 values (in brackets) are included in LOI

	55-5B2	562A	655	556A	19A-7B	19A-6D	64-1C4	17-15	53-3C1	554	583	54-4F1
SiO_2	74.3	70.5	74.3	69.6	72.8	67.4	66.3	65.3	73.7	71.0	66.0	70.8
Al_2O_3	13.9	15.2	14.4	14.8	14.4	15.5	16.6	15.3	13.5	15.1	15.5	15.5
CaO	0.05	0.60	0.99	0.93	1.94	0.72	0.87	3.21	0.28	0.44	1.17	0.76
MgO	0.44	0.57	0.45	1.10	0.62	1.05	1.08	2.15	0.83	0.96	1.13	0.99
Na_2O	0.12	5.84	4.78	5.31	6.02	5.94	6.22	2.97	3.76	5.21	4.92	3.08
K_2O	8.29	4.22	2.12	3.99	1.12	3.89	4.30	2.87	4.00	3.74	4.65	5.22
Fe_2O_3	1.32	2.09	1.29	2.50	1.38	3.36	2.74	5.05	2.55	1.91	4.32	1.81
MnO	0.01	0.06	0.03	0.05	0.03	0.10	0.05	0.14	0.03	0.04	0.08	0.06
TiO_2	0.28	0.37	0.04	0.40	0.21	0.56	0.45	0.47	0.29	0.39	0.89	0.24
P_2O_5	0.02	0.09	0.08	0.13	0.03	0.16	0.13	0.15	0.05	0.09	0.23	0.05
BaO	0.17	0.17	0.06	0.14	0.02	0.13	0.10	0.13	0.09	0.14	0.11	0.13
LOI	1.16	0.39	1.54	0.77	0.85	1.00	0.93	1.93	1.08	1.08	0.93	1.39
CO_2	(0.1)	(0.1)	(0.7)	(0.1)	(0.1)	(0.2)	(0.1)	(0.2)	(0.1)	(0.1)	(0.1)	(0.1)
TOTAL	100.2	100.2	100.1	99.8	99.5	99.9	99.8	99.7	100.2	100.2	100.0	100.1
Rb	380	100	60	120	60	120	90	80	130	100	120	220
Sr	150	190	120	200	750	220	290	400	60	160	140	250
Y	11	26	11	22	20	n.d.	25	25	44	25	43	21
Zr	260	270	20	200	140	220	220	140	280	270	350	140
Nb	14	12	14	11	9	n.d.	11	9	16	13	14	12
Ba	1468	1460	515	1225	175	350	875	1140	788	1225	963	1140
Ni	2	<2	<2	<2	<2	n.d.	4	5	3	<2	4	<2
Cr	5	10	10	10	10	10	10	10	10	10	10	10
V	200	20	<10	40	10	50	n.d.	80	10	n.d.	n.d.	n.d.
Ga	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Sc	5.2	4.3	1.0	5.4	2.9	9.4	n.d.	12.1	9.9	n.d.	n.d.	n.d.
Th	30.0	13.0	3.1	11.0	13.0	9.6	n.d.	10.0	10.0	n.d.	n.d.	n.d.
La	30.0	50	18.2	40.4	38.0	36.5	n.d.	30.5	28.3	n.d.	n.d.	n.d.
Ce	49	77	32	62	57	56	n.d.	54	72	n.d.	n.d.	n.d.
Nd	14.0	30.0	14.0	22.0	16.0	23.0	n.d.	19	24	n.d.	n.d.	n.d.
Sm	3.7	5.5	3.6	4.4	3.0	4.9	n.d.	4.7	5.7	n.d.	n.d.	n.d.
Eu	0.3	0.4	0.3	0.4	0.2	0.7	n.d.	0.7	0.4	n.d.	n.d.	n.d.
Yb	2.1	3.1	0.5	2.3	1.9	2.5	n.d.	2.9	4.4	n.d.	n.d.	n.d.
Lu	0.47	0.54	0.09	0.42	0.36	0.51	n.d.	0.47	0.78	n.d.	n.d.	n.d.
Hf	9	9	2	6	5	7	n.d.	5	10	n.d.	n.d.	n.d.
Ta	<1	<1	1	1	1	1	n.d.	1	1	n.d.	n.d.	n.d.

n.d. = not determined

The modern subduction-related basalts commonly contain anomalously low Nb contents. This is attributed by Saunders and Tarney (1984) to the enhanced stability of Nb-bearing phases (such as rutile and ilmenite) in the mantle wedge source rock during melting under conditions of high $p(H_2O)$ or $p(O_2)$. Oxidizing fluids arise as a result of the dehydration of the subducted slab in an island arc environment (Fig. 20). These fluids are often LIL-enriched and may be responsible for the mantle metasomatic event associated with the genesis of these basalts. Thus, it is postulated that the mantle source for the basalts was contaminated with LIL-enriched fluid prior to or during melting.

Felsic rocks

The two most common models proposed for the origin of felsic rocks in volcanic arc environments are fractionation of a more mafic parent and crustal anatexis. The low concentrations of Zr, Nb (Table 3) and REE in the felsic volcanic rocks (Fig. 17e), the bimodality of Keppoch volcanism with respect to SiO_2 and the relative abundance of mafic to felsic rocks suggest it is unlikely that the felsic volcanic rocks were derived from either the basalts or the basaltic andesites by fractionation.

It is more likely that they were derived by partial melting of the crust associated with the high heat flow due to the rising mafic magma (e.g. Barker, 1981). This is difficult to evaluate because no information on the nature of the source region is available.

Summary

The stratigraphy and depositional environment of the Georgeville Group in the Antigonish Highlands is consistent with the progressive opening of a sedimentary basin (see the section on Stratigraphy above). The geochemistry of the volcanic rocks supports this model and suggests that three sources of magma were present at the incipient stages of basin

Table 3c. (cont.)

	54-4E3	54-4E2	405B	17-10	24-22D	17-11	53-3A	56-1B	56-3D	54-4F2	657	405A
SiO_2	74.4	76.5	68.4	72.4	70.3	69.7	72.9	67.3	68.4	68.5	75.6	62.7
Al_2O_3	13.1	12.3	15.2	13.6	14.9	15.1	14.0	16.6	15.7	16.8	13.7	19.1
CaO	0.99	0.61	1.50	0.54	0.96	0.39	0.38	1.88	1.26	0.69	0.83	0.84
MgO	0.58	0.59	0.81	0.98	0.47	0.90	0.61	0.54	0.62	1.02	0.38	2.42
Na_2O	3.17	0.62	3.55	2.77	6.25	3.38	3.40	2.96	4.81	3.08	4.46	3.72
K_2O	4.40	6.33	5.01	5.70	2.38	6.66	4.64	6.74	4.27	5.99	2.28	4.08
Fe_2O_3	1.45	1.26	3.16	1.61	2.95	2.01	2.56	1.93	2.75	1.81	1.14	3.55
MnO	0.04	0.03	0.06	0.17	0.03	0.07	0.04	0.06	0.07	0.06	0.02	0.09
TiO_2	0.20	0.19	0.43	0.28	0.11	0.36	0.31	0.26	0.55	0.28	0.02	0.55
P_2O_5	0.05	0.06	0.14	0.08	0.11	0.09	0.07	0.09	0.11	0.07	0.08	0.17
BaO	0.10	0.11	0.17	0.19	0.10	0.26	0.12	0.12	0.14	0.12	0.06	0.13
LOI	1.23	1.23	1.16	1.62	1.31	1.31	1.00	1.23	1.08	1.54	1.62	2.85
CO_2	(0.1)	(0.1)	(0.1)	(0.4)	(0.1)	(0.2)	(0.1)	(0.1)	(0.1)	(0.1)	(0.6)	(0.1)
TOTAL	99.8	99.9	99.7	100.0	99.9	100.3	100.1	99.8	99.8	100.0	100.2	100.3
Rb	180	290	160	190	80	180	160	290	150	250	80	180
Sr	240	130	380	140	300	180	120	330	280	300	110	150
Y	19	9	20	20	5	21	40	18	36	18	8	38
Zr	110	100	230	160	90	200	320	150	240	150	20	290
Nb	9	13	9	8	2	6	15	15	15	13	13	12
Ba	875	960	1460	1660	875	2275	1050	1050	520	520	520	1140
Ni	<2	<2	<2	2	5	4	<2	4	2	2	<2	4
Cr	10	10	5	5	10	10	10	5	10	10	10	10
V	20	20	n.d.	20	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	<10	n.d.
Ga	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Sc	3.1	3.8	n.d.	3.3	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	1.0	n.d.
Th	28	26	n.d.	10	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	2.7	n.d.
La	48.2	38.2	n.d.	39.4	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	9.2	n.d.
Ce	65	53	n.d.	59	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	16	n.d.
Nd	19	15	n.d.	17	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	8	n.d.
Sm	3.7	3.0	n.d.	3.4	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	2.4	n.d.
Eu	0.4	<0.2	n.d.	0.2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.2	n.d.
Yb	1.6	1.3	n.d.	1.8	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.5	n.d.
Lu	0.32	0.32	n.d.	0.35	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.1	n.d.
Hf	5	4	n.d.	6	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	2	n.d.
Ta	1	1	n.d.	<1	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	2	n.d.

n.d. = not determined

development: (i) a calc alkalic (with some tholeiitic tendencies) basaltic andesite; (ii) a tholeiitic-alkalic basalt that produces basalts at different stratigraphic levels and may be genetically associated with subduction; and (iii) a felsic magma that has volcanic arc characteristics.

The Keppoch Formation is probably partly contemporaneous with the Chisholm Brook Formation. The geochemistry and the lithological association suggests that these formations represent magmatism in the incipient stages of the development of an ensialic marine basin within a volcanic arc. The presence of a calc-alkalic component in these rocks suggests proximity to a subduction zone in which the mantle wedge overlying the subduction zone was contaminated by LIL-enriched fluids or by calc-alkalic magma.

The stratigraphic context indicates that the basalts of the Keppoch Formation are continental whereas the basalts of the Clydesdale Formation are deposited in a marine basin. The linear magnetic anomaly may be attributed to the stretching of the lithosphere over the "hot-spot" during basin development or to magmatism concentrated on a major line of weakness. These basalts have anomalously low Nb contents, which is typical of basalts genetically associated with subduction (Saunders and Tarney, 1984; Pearce et al., 1984b).

It is suggested that the calc-alkalic geochemical signature of the basaltic andesites, the within-plate continental character of the basalts and the volcanic arc compositions of the felsic rocks probably represent a transitional tectonic setting (orogenic to ocean floor) characteristic of the early stages of rifting in a volcanic arc environment. According to Saunders and Tarney (1984), penecontemporaneous calc-alkalic and tholeiitic geochemical signatures are best developed in narrow ensialic basins. This environment is consistent with the stratigraphic succession for the Georgeville Group in which a subaerially deposited volcanic-dominated sequence is overlain by a thick succession of turbidites. The bimodal nature of the volcanism with respect to SiO_2 is also typical of a rifting environment (e.g. Martin and Piwinski, 1972). Geochronological data which indicate a limited temporal gap between deposition, deformation, and intrusion of the Georgeville Group suggest a strike-slip origin for the basin.

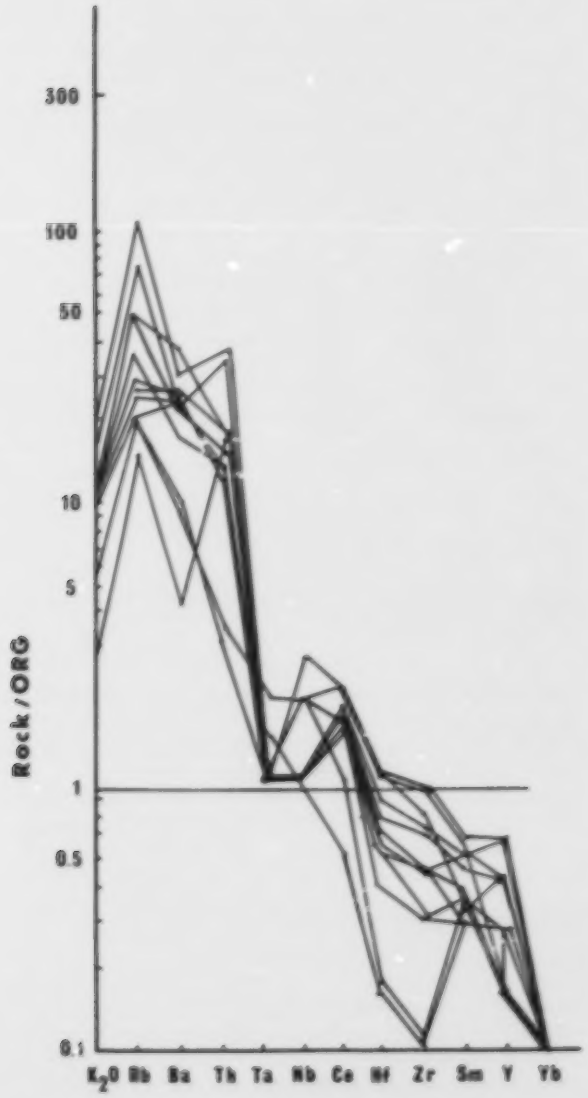


Figure 23. Felsic volcanic rocks of the Keppoch Formation normalized to ORG (see Pearce et al., 1984a,b; Saunders and Tarney, 1984).

Discussion

The Avalon Zone (Williams, 1978, 1979) is a composite terrane produced by the amalgamation of distinct Precambrian terranes during the latest Precambrian Avalonian-Cadomian Orogeny (Keppie, 1985). At present, there is insufficient geochronology to document whether these

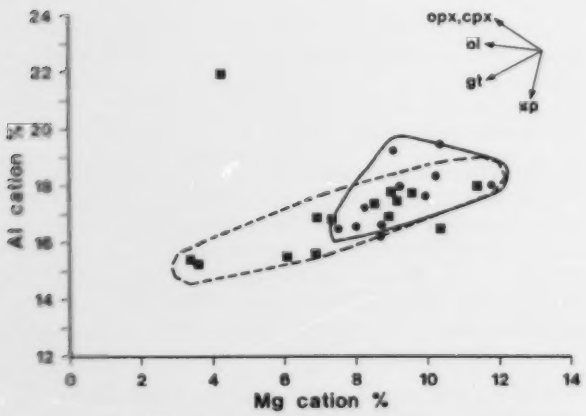


Figure 24. Al versus Mg cation plot for the basaltic rocks of the Clydesdale and Keppoch formations. Symbols as in Figure 15. opx-orthopyroxene, cpx-clinopyroxene, ol-olivine, gt-garnet, sp-sphene. Solid line encloses Clydesdale analyses. Dashed line encloses Keppoch analyses.

differences are between time-equivalent units (thereby defining distinct terranes), or whether they reflect units of different ages (which could be either distinct terranes, or different stratigraphic levels within one terrane). Thus, at this stage, only the gross spatial distributions can be compared.

The late Precambrian rocks in Nova Scotia show two distinct associations. In southern Cape Breton Island, the Fourchu Group is made up of subaerial to shallow submarine volcanic rocks of calc-alkalic affinity (Keppie et al., 1979; J. Dostal, pers. comm., 1988). On the other hand, in the Antigonish Highlands, the Georgeville Group is predominantly made up of turbidites and three distinct volcanic suites (calc-alkalic basaltic andesites, tholeiitic-alkalic basalts and felsic volcanic rocks). The late Precambrian rocks of the Cobequid Highlands show some broad chemical similarities with the Keppoch Formation (Pe-Piper and Piper, 1987, 1989; Murphy et al., in press).

The volcanism represented by the Fourchu Group has been attributed to an ensialic island arc tectonic environment above a northwesterly dipping subduction zone (Keppie et al., 1979). Thus Keppie (1982) and Keppie and Dostal (in press) interpret the Antigonish Highlands as a marine basin behind the Fourchu arc. Although a temporal relationship with the Fourchu Group has yet to be established, this model is supported by the following observations in the Antigonish Highlands:

1) the stratigraphic sequence records the progressive development and infilling of a sedimentary basin;

2) the thickness of the clastic sediments of the Georgeville Group suggests that this basin was close to land;

3) the geochemistry of the basaltic andesites shows calc-alkalic affinities, the felsic rocks have volcanic arc compositions and the tholeiitic-alkalic basalts have a possible supra-subduction zone chemistry.

The contamination of basaltic andesites and basalts with oxidizing fluids is attributed to dehydration of the subducting slab.

Further evaluation of the tectonic history of the area awaits age control, but the geological history of the Antigonish Highlands places further constraints on the evolution of the late Precambrian history of the Avalon Terrane in Nova Scotia, and the Appalachians as a whole.

The relationship between the late Precambrian rocks of the Georgeville Group and those in central and northern Cape Breton Island is uncertain. The gneissic basement of the George River Group has been interpreted as basement to the Fourchu Group (Helmstaedt and Tella, 1973; Keppie, 1982, 1990; Currie, 1982; O'Brien et al., 1983) or as part of a distinct terrane (Barr and Raeside, 1986, 1989). Further work may resolve this problem.

The late Precambrian rocks of southern New Brunswick appear to have gross similarities to those of Cape Breton Island (Giles and Ruitenberg, 1977; Currie, 1986). They consist of a gneissic rocks (Brookville Gneiss), a shallow marine unit (Greenhead Group--equivalent to the George River Group of Nova Scotia) and late Precambrian volcanic rocks (Coldbrook Group). The Coldbrook Group is interpreted to represent a continental rift setting similar to those in Newfoundland (Giles and Ruitenberg, 1977). Geochemical data of the Coldbrook Group is similar to that of the Keppoch Formation (Currie and Eby, 1990).

The late Precambrian rocks of the Avalon Zone in Newfoundland consist mainly of subaerial to shallow marine volcanic rocks overlain in turn by flyschoid sediments (Conception and Connecting Point groups) and molassic rocks (King, 1980) passing with or without unconformity into Cambro-Ordovician rocks. These volcanic rocks range in age from about 630 to 570 Ma (Krogh et al., 1983). They are generally bimodal and alkalic-tholeiitic, and were extruded in a continental within-plate environment (Strong et al., 1978; Cameron, 1986).

O'Brien et al. (1983) propose that the tectonic evolution of the Avalon Terrane is transitional between that of an intra-plate and continental margin and is dominated by repeated rifting events from 800 to 570 Ma. These authors propose a correlation with the late Precambrian Pan-African belts (compare Kroner, 1979). This is supported by recent paleomagnetic data (e.g. Johnson and Van der Voo, 1986) which suggest a genetic link between the Fourchu Group and the African craton in the late Precambrian. The rifting events are thought to represent repeated attempts of the African continent to break apart in the upper Proterozoic. If this is the case, then the presence of subduction-related magmatism in the Avalon Zone of Nova Scotia suggests that these rifts, in some cases, may have formed from relatively large ocean basins and that the subduction-related magmatism and the development of inter-arc basins is associated with the destruction of these ocean basins. However, the lack of abundant andesitic-dacitic rocks suggests that although the basin was wide enough that its destruction could generate subduction-related volcanism, it was not wide enough to generate a mature island arc. Alternatively, the major island arc system may have been an outboard of the African craton, and the volcanic rocks of Nova Scotia may thus represent local rifts associated with subduction.

Murphy and Nance (1989) interpret the inception and termination of the Avalonian orogenic cycle to be related to the amalgamation and breakup of a late Precambrian supercontinent. Synorogenic volcano-sedimetnary and plutonic rocks are related to subduction along the margin of the supercontinent after its amalgamation. Breakup of the supercontinent led to termination of subduction and its replacement by tranform activity.

Structure and metamorphism

Introduction

Interpretation of the structural history of the Georgeville Group is hindered by the lack of outcrop in the inland areas of the highlands and by the lack of small-scale structures. Most of the descriptions and interpretations below are based on regional mapping. Structural styles and deformational history in the Georgeville Group vary considerably from north to south within the highlands. In general, north of the

Browns Mountain and Marshy Hope faults (i.e. the Georgeville, Maple Ridge and Clydesdale blocks, Fig. 3) the Georgeville Group is polydeformed with several generations of folding; to the south of these faults (the Keppoch Block), the Georgeville Group is only moderately deformed, folding is supratenuous (i.e. drape folds) and the structural style is dominated by faulting.

Deformation in the Georgeville, Maple Ridge and Clydesdale blocks

D_1 structures (F_1, S_1)

The earliest tectonic structures observed in the late Precambrian Georgeville Group are tight-isoclinal folds and a sporadically developed axial planar cleavage (S_1) which is commonly parallel to bedding (S_0). The S_1 cleavage is commonly defined by sericite, chlorite and minor biotite and is well developed in the mudstones (where in some localities it is penetrative), but is weakly developed in the greywackes and volcanic rocks. The rarely exposed minor F_1 folds generally occur at an outcrop scale and conform to cylindrical class 1C of Ramsay (1967). The presence of major F_1 folds is commonly indicated by opposing facing directions in sections with uniform bedding attitudes. However, in areas of poor exposure, the positioning of F_1 hinges is more tenuous. For example, on South Rights River, areas of high magnetic signature are assumed to be underlain by mafic volcanic rocks of the Clydesdale Formation. This assumption, together with stratigraphical considerations, was used to define an F_1 hinge about 6 km west of Clydesdale (see Map 1749A, in pocket). It is conceivable, however, that this area may be underlain by mafic intrusive rocks.

Other F_1 structures are well exposed. For example, a major F_1 anticlinal structure that can be mapped over a length of 30 km occurs in the Maple Ridge block from the Maple Ridge area to Eigg Mountain and Vameys Brook (see Map 1749A). The structure is truncated by a fault near Vameys Brook and extrapolation is somewhat obscured by unconformably overlying Silurian strata. However, a major F_1 anticlinal structure is interpreted in the Bears Brook area, where opposing facing directions occur on shallowly dipping monoclinal successions. The best indications for this structure can be seen in outcrops exposed on a path that follows the western bank of Bears Brook. The Maple Ridge Formation occurs in the core of the fold and the James River Formation occurs at the flanks. The hinge surface trace varies from north-northeast to northeast trending. It is defined by localities with a high angle between S_0 (bedding) and S_1 (cleavage) (Fig. 25). These localities are best exposed in open fields east of Vamey River. Cleavage in the hinge zone is more strongly developed making bedding less visible than on the fold limbs. S_0/S_1 intersections in the hinge region indicate that the fold plunges shallowly or moderately to the northeast. A lower hemisphere equal-area projection (Fig. 26 and 27) shows the distribution of poles to bedding and cleavage on the northern and southern limbs of this F_1 structure in the Maple Ridge, Eigg Mountain and Bears Brook areas. The general parallelism between S_0 and S_1 indicates that the F_1 fold is tight to isoclinal. The attitude of the cleavage is generally steep to vertical in the Maple Ridge and Eigg Mountain areas indicating that the fold is generally upright (Fig. 26 and 27a). In the Bears Brook area, the cleavage and bedding are shallow to moderately dipping (Fig. 27b). The change in attitude of the F_1 structure is attributed to later deformation. F_1 folds may also be observed in the outcrop patterns on the geological map (Map 1749A), especially in the Clydesdale block (Fig. 28).

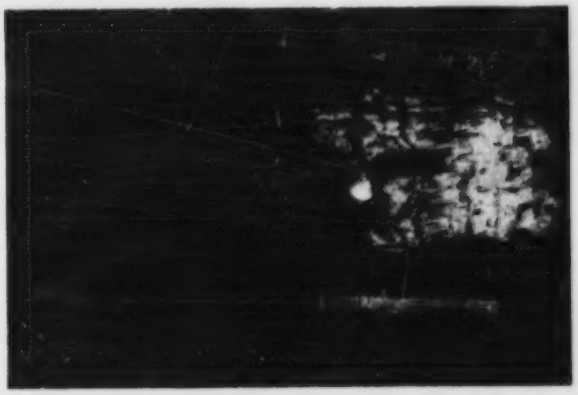


Figure 25. S_0/S_1 relationship displayed in the hinge of a regional F_1 fold structure, Maple Ridge Formation. Bedding and cleavage are perpendicular. The cleavage strikes northeast (parallel to the hammer). GSC 204717B.

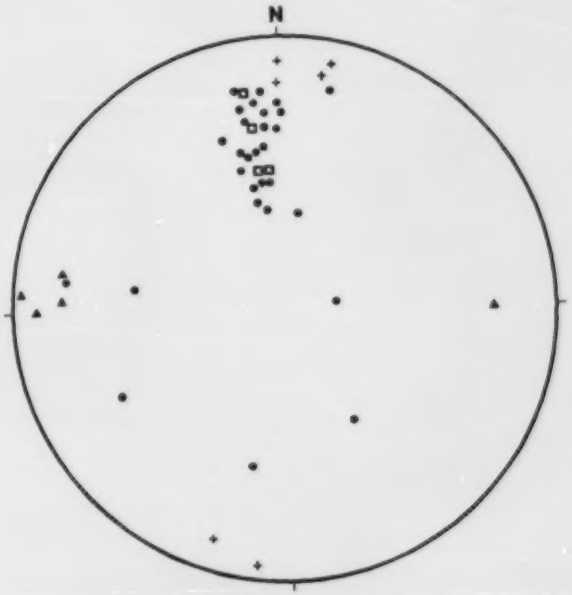


Figure 26. D_1 structural data from the Maple Ridge block, Maple ridge area. Filled circles-poles to bedding, open squares-poles to S_1 cleavage, filled triangles-poles to S_2 fracture cleavage, crosses-poles to S_3 fracture cleavage.

It is clear from these major folds that most of the Georgeville Group, including the volcanic rocks, was involved in this deformation. This feature indicates that the major F_1 folds are tectonic rather than soft-sediment in origin. The sense of vergence and original attitude of the F_1 folds may be deduced from the unfolding of later structures (Fig. 28) in the Brierly Brook area. Although outcrop is only moderate in this area, the overall distribution of the lithologies is consistent with a mushroom interference pattern between F_1 and F_3 structures. This indicates that the F_1 folds were easterly vergent and probably had north-trending fold axes prior to F_3. However, because F_2 folds are approximately coaxial with F_1 and are open (see below), the F_2 folding event probably did not affect the vergence or the orientation of the F_1 folds. This implies that these orientations approximate the original orientations of the F_1 folds. The regional extent of the major recumbent F_1 folds is uncertain. They appear to be confined within individual fault-bounded blocks. For example, regional F_1 structures have not been found in the Keppoch block.

The presence of sericite, chlorite and minor biotite defining the S_1 axial planar cleavage in the pelitic rocks indicates that greenschist facies metamorphic conditions accompanied the first set of structures. Albite, epidote, quartz and calcite may also be present in these rocks. Limestones in the Morar Brook Formation distal from plutonic bodies contain calcite-quartz-chlorite-epidote-bearing assemblages, also indicative of greenschist facies metamorphism. Metamorphic assemblages in the basalts and basaltic andesites defined by albite-epidote-chlorite ± actinolite ± calcite ± quartz are also consistent with greenschist facies metamorphism. However, it is possible that some of these minerals were formed during seafloor alteration prior to regional metamorphism.

D_2 structures (F_2, S_2, L_2)

D_2 structures are readily distinguished from D_1 structures because they fold the S_1 cleavage. They are best developed in the eastern Maple Ridge and Clydesdale blocks (Map 1749A; Fig. 28). In contrast to the adjoining area in the eastern highlands, these folds result in local modification of the attitudes of structures rather than seriously affecting the regional distribution of formations. Lower hemisphere equal-area projections of bedding and cleavage data from F_2 folds associated with the D_2 deformation are shown in Figure 29a,b. F_2 folds are northeast trending (Maple Ridge block) to north trending (Clydesdale block) and are generally plunging upright open folds. They have a weakly developed axial planar fracture cleavage, broad hinge zones and a wavelength of up to 5 km. Congruous minor folds associated with F_2 are rare.

Regional F_2 hinge zones have not been defined with any certainty in the western part of the Maple Ridge block although one F_2 hinge has been tentatively interpreted (Map 1749A). The lack of regional F_2 fold closures in this area may reflect an east to west structural heterogeneity, or more probably indicates that the area is located on the limb of a major F_2 fold.

The change in fold trend from north-south in the Clydesdale block, through north-south to northeast-southwest in the Maple Ridge block to northeast-southwest in the Georgeville block (Map 1749A; Fig. 3 and 28) is inferred to be the result of either later F_3 folding or to dextral transcurrent motion on the Hollow Fault. Bedding/S_2 cleavage intersection lineations define a girdle in the plane of the cleavage (Fig. 29a,b). This pattern cannot be attributed to later F_3 folding since it occurs on one limb of an F_3 structure. It could be due

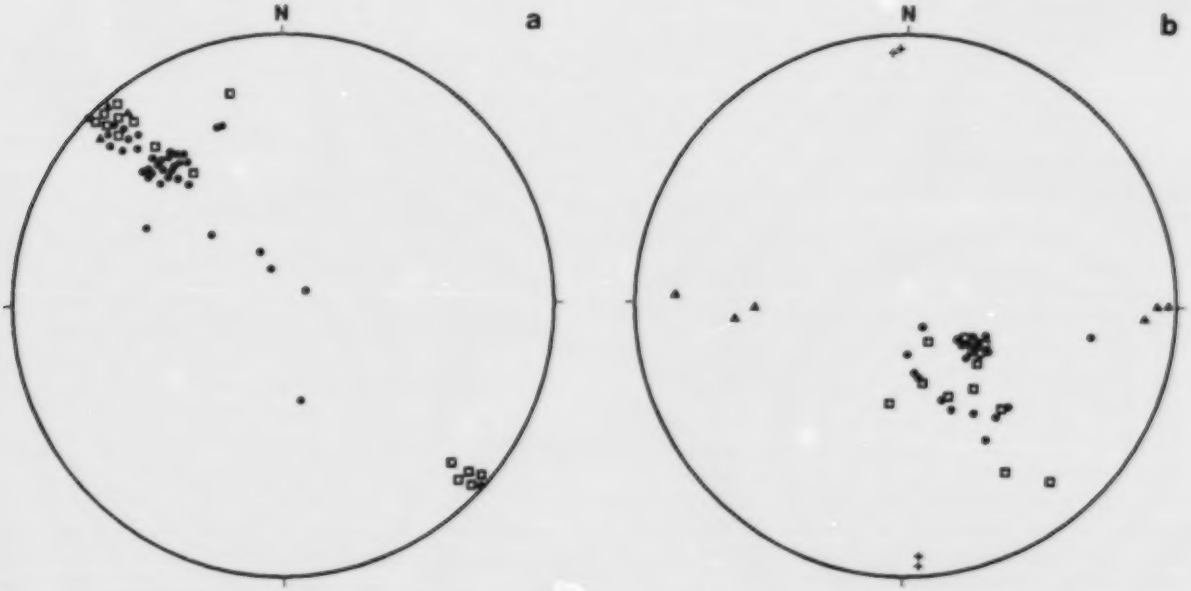


Figure 27. D_1 structural data from (a) Eigg Mountain and (b) Bears Brook areas, Maple Ridge block. Symbols as in Figure 26.

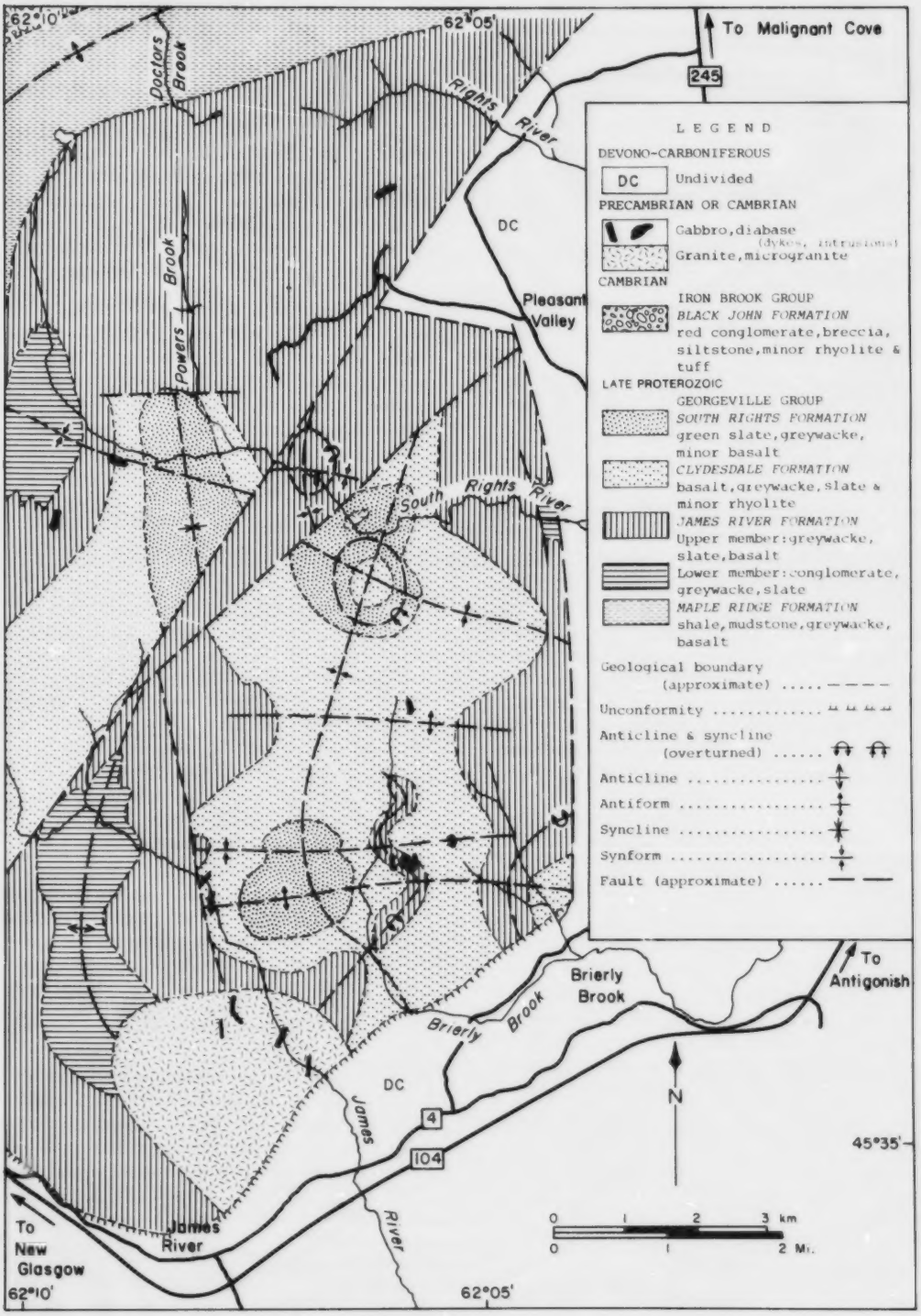


Figure 28. Summary of the geology and structure of the Clydesdale block.

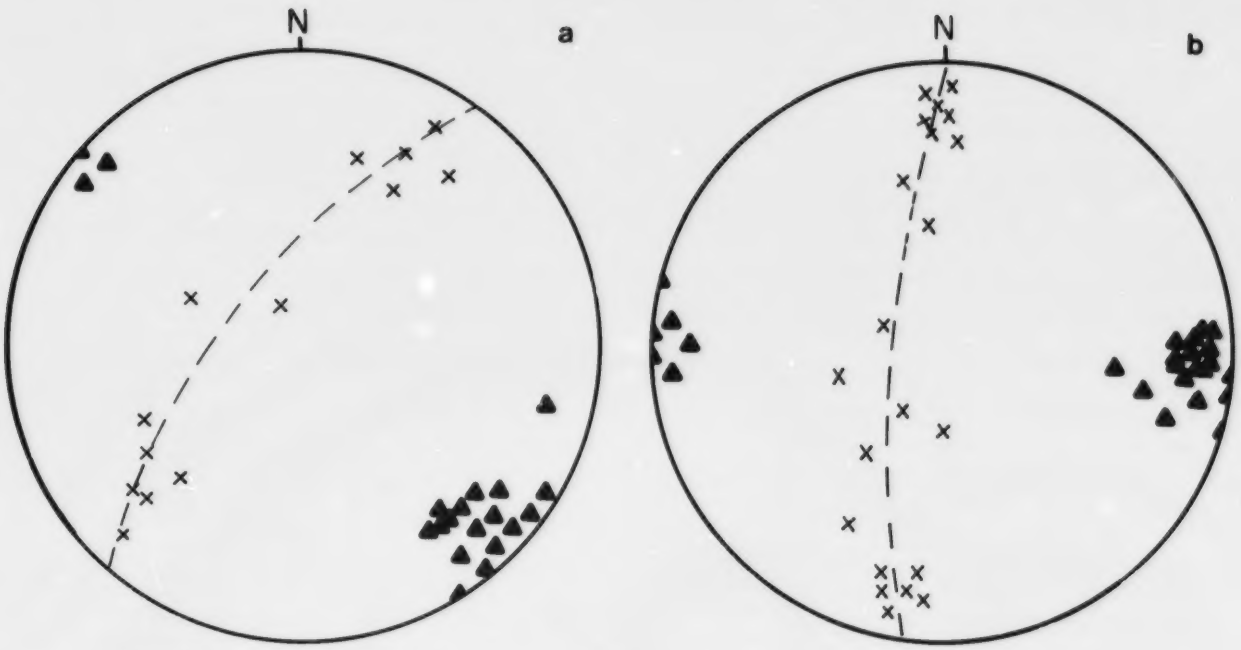


Figure 29. D_2 structural data from the (**a**) Maple Ridge and (**b**) Clydesdale blocks. Triangles-poles to S_2; crosses-S_0S_2 intersection.

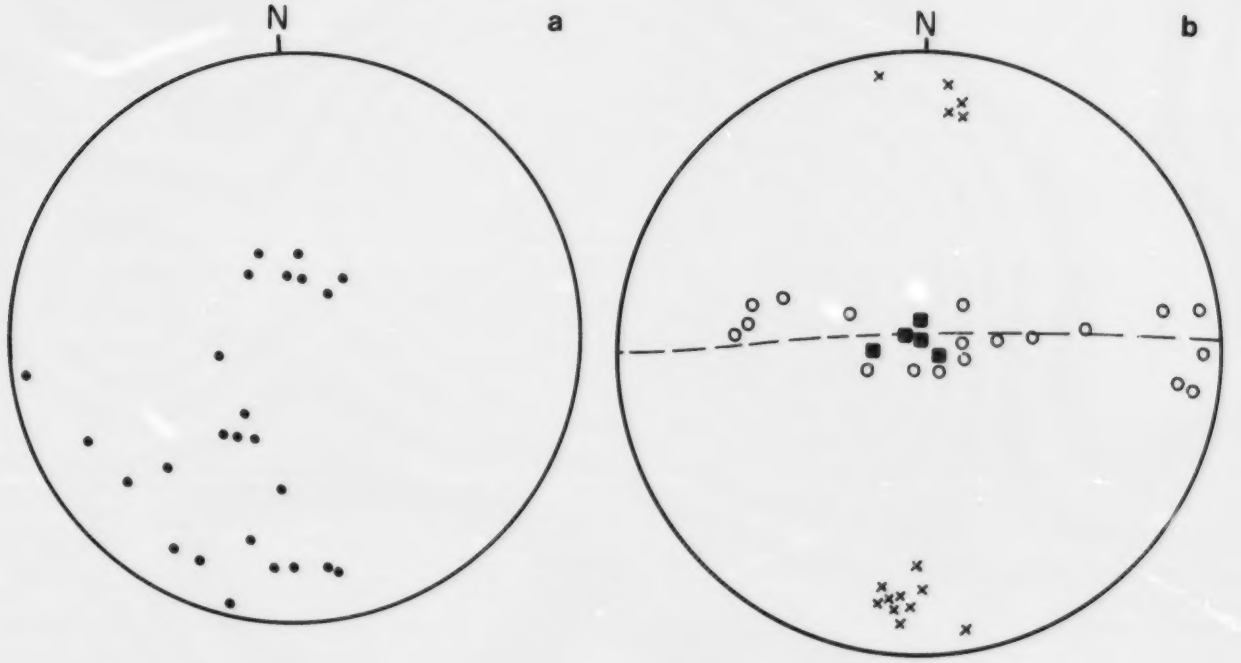


Figure 30. (**a**) D_1 and (**b**) D_3 structural data from the Clydesdale block. Filled circles-poles to bedding, crosses-poles to S_3, circles-S_3S_0 intersections, filled squares-S_3S_2 intersections.

to one or more of: (i) the superposition of F_2 on earlier F_1 folds; (ii) inhomogeneous strain in the S_2 cleavage planes during or after folding (e.g. by motion on the Hollow Fault); or (iii) en echelon style of folding.

D_3 structures (F_3, S_3, L_3)

East-west upright, open to closed folds with broad hinge zones and a wavelength of about 5 km occur in both the Maple Ridge and Clydesdale blocks. Although regional F_3 folds are common, minor congruous folds are rare. The large east-west fold in the Maple Ridge block produces a swing in the trend of F_1 and F_2 folds (Map 1749A). A fracture cleavage S_3 is axial planar to these folds.

In the Clydesdale block, bedding defines a conical pattern of varying apical angles (Fig. 30a) about F_2 structures. The bedding/S_3 cleavage intersection lineation define a planar distribution in the S_3 cleavage (Fig. 30b) due to superposition of the S_3 cleavage upon previously folded bedding. In contrast, S_3/S_2 intersection lineations are concentrated around a vertical attitude (Fig. 30b). The lack of any recrystallization in the S_3 fracture cleavage indicates the low subgreenschist grade of metamorphism at this time. In the western Maple Ridge block, F_3 folding results in the local modification of the attitudes of bedding and S_1 cleavage. F_3 folding does not result in a significant redistribution of formations.

Deformation in the Keppoch block

In contrast to the Georgeville, Maple Ridge and Clydesdale blocks, no regional folds can be identified in the Keppoch block. The turbidites in the Keppoch block predominantly consist of monoclinal sequences that dip and young away from a central core of Keppoch volcanic rocks (Map 1749A). Thus, in the northern and southern parts of the block, the turbidites strike east-west and dip (respectively) to the north and south. In the eastern and western parts of the block they strike north-south and dip to the east and west respectively. Although way-up evidence is lacking in some areas, the available evidence indicates that these rocks are predominantly right-way-up, suggesting that regional F_1 recumbent folds do not occur. A single example of a local F_1 fold was found on Baxter Brook.

S_0/S_1 relationships in the greywackes in the Keppoch block also contrast with the northern highlands. Cleavage is generally more steeply dipping than bedding (Map 1749A). Cleavage-bedding intersection lineations are subhorizontal and are west trending in the north and south of the block, with north-trending lineations in the east and west of the block. These structural relationships indicate that the folds are predominantly upright in this domain. This interpretation is consistent with the facing evidence mentioned above.

Deformation in the Keppoch Formation is difficult to evaluate owing to probable complex primary distribution of the volcanic rocks (due to subaerial deposition), lack of way-up criteria, penetrative foliation and abundant faults. There is evidence from the map pattern that the Keppoch Formation may have controlled structural trends in younger rocks. If this is correct, then the Keppoch Formation behaved in a competent manner during the deformation of the younger rocks. Bedding in the Keppoch Formation generally trends east-west in the north and south of the area and north in the eastern parts of the area (Map 1749A). Strikes in younger strata (Precambrian to Carboniferous) generally parallel these trends (Map 1749A). This observation generally holds whether the contacts are unconformable or faulted, suggesting that primary distribution of the relatively competent Keppoch volcanic rocks had a significant effect on determining younger structures. These effects die out away from the contacts.

The Keppoch Formation rarely displays a tectonic fabric and may have responded in a brittle fashion and/or acted as a rigid mass during deformation of the Georgeville Group during subsequent deformations. The map pattern of the Keppoch Formation gives no information as to whether or not it is folded on a regional scale.

Thus, there is no compelling evidence for regional folding within the Keppoch block. It is likely, since much of the Keppoch Formation was deposited subaerially, that the distribution of units was governed by paleotopography. Therefore the present distribution may be mainly governed by primary considerations. If regional folding did occur in the Keppoch block, it appears to have been supratenuous, i.e. antiformal fold hinges are located on paleotopographic highs and limbs are on paleoslopes. If this is true, then tectonic folding may have resulted in a tightening of the primary distribution. Further evidence for supratenuous folding is provided by the thinly laminated mudstones ("Georgeville undivided" on Map 1749A) that drape the southwestern and southern highlands (Fig. 3; Map 1749A) in a series of monoclinal successions that consistently dip away from the Keppoch Formation throughout the southern highlands.

Age of deformation

The following relationships constrain the age of D_1 and D_2 deformational events

1) In the Doctors Brook area, the Maple Ridge Formation is unconformably overlain by Lower Cambrian rocks (Fig. 3; Map 1749A). Although no angular discordance is visible at the exposed contact, stratigraphical considerations (i.e. juxtaposition of the Maple Ridge Formation which occurs toward the base of the Georgeville Group with Lower Cambrian rocks) indicate that this is a major regional unconformity.

2) The Greendale Complex (Fig. 3) dated at ca. 620-608 Ma (Ar^{40}/Ar^{39} on hornblende; Keppie et al., 1990) cuts across S_1 fabrics and apparently truncates F_2 folds in the Georgeville block give an upper age limit.

3) The presence of 613 ± 5 Ma detrital zircons in the Livingstone Cove conglomerates give a lower age limit (Keppie and Krogh, 1990).

4) Thin variably oriented pegmatitic and granitic veins associated with an adjacent stock of alaskite intrude the Morar Brook Formation of the Georgeville Group, apparently truncate F_2 structures and are undeformed

adjacent to isoclinal F_1 fold hinges, giving a minimum age for F_1 and F_2 deformation. The alaskite has been dated at 604 ± 14 Ma (K-Ar on muscovite; K. Wanless, pers. comm., 1980).

5) F_1 folds cannot be found in Cambrian rocks. Although Cambrian rocks have been deformed by recumbent structures, these structures are localized to the centre of the Cambrian basin and die out toward the margins where merely tilted Cambrian rocks rest on the Georgeville Group (see section on Structure and Metamorphism for Cambro-Ordovician Rocks below).

6) Pebbles of greywacke and mudstone, which are similar to the Georgeville Group and have a penetrative fabric, are found in Lower Cambrian conglomerates.

The data, taken together indicate a ca. 615-610 Ma age for D_1 and D_2 deformation. The age of the D_3 structures is uncertain because no definitive crosscutting relationships with late Precambrian plutons or lower Paleozoic rocks are observed. The structures are truncated by faults (e.g. the Browns Mountain Fault) of which the latest movement is unknown (see Keppie, 1983).

Synthesis of Precambrian deformation

Origin of the structures

The Precambrian deformation in the Georgeville Group probably represents the closure of the marine basin. This is supported by the ca. 615-610 Ma age of the structures and by the unconformable relationships of the Georgeville Group with the overlying subaerially deposited Cambrian rocks. The fragmentary knowledge of the earliest recumbent F_1 folds and the lack of kinematic-indicating fabrics in the Georgeville Group make it difficult to present a rigorous kinematic model for the genesis of these rocks. One of the principal difficulties is the lack of fabrics that would indicate the direction of emplacement of the F_1 nappes. The localized nature of the folding suggests that the folds may represent a positive flower structure. This is consistent with a strike-slip origin for deposition and deformation of the basin indicated by the limited temporal gap between deposition and deformation.

F_2 folds are coaxial with F_1 folds, have a north trend and probably represent a second "roll-over" phase of folding as emplacement of the thrusts proceeded. If this is correct, then D_1 and D_2 are continuous. This interpretation is supported by the above geochronological data. Synchronous folding and transcurrent movements on northeast-trending faults may be responsible for the variation in attitude of the fold axes of the F_2 folds (Fig. 29). Alternatively, if the F_2 folds represent a distinct deformational pulse rather than a progression of thrusting, they may have been produced by east-west compression which would result in dextral movement on northeast-trending faults. If this hypothesis is true, in more intensely sheared areas adjacent to those faults the north-south folds would swing toward a northeast-southwest trend, as is observed (Map 1749A). This swing defines an east-west hinge in the Maple Ridge block. This mechanism probably contributed to the trend variations which were also produced by later F_3 folding.

The age and origin of the F_3 folds is problematic. The swing in the orientation of the F_2 folds in the Maple Ridge block may be attributed to dextral motion on the Hollow Fault. Such a motion has been proposed for the Hollow Fault in the Late Precambrian (Murphy and Hynes, 1990) Cambrian (Murphy et al., 1985) and the Devono-Carboniferous (e.g. Keppie, 1983; Yeo and Ruixiang, 1986). It is therefore conceivable that the F_3 structure in the Maple Ridge block may be as young as Carboniferous in age.

In contrast, the east-west F_3 folds in the Clydesdale block do not affect the orientation of the axial planes of F_2. It is not certain if they are the same generation of folds as the large east-west fold in the Maple Ridge block. If they were associated with north-south compression, this would have produced sinistral transcurrent movements of the major northeast-southwest faults. Consideration of the whole Precambrian structural history allows that F_1 and F_2 could be the result of one period of progressive deformation. During the early stages of this deformation recumbent nappes were produced followed progressively by north-south refolding of the nappes. Of course, separate deformational pulses are another possibility that cannot be eliminated.

The map pattern clearly demonstrates that volcanic rocks were involved in the deformation, arguing strongly for a significant hard-rock (tectonic) component to the deformation. Furthermore, if the deformation had a significant soft-sediment component one would expect slumping to be most intense at volcanic-sediment contacts. This is not observed. For example, near Bears Brook, internally deformed beds up to 50 cm in thickness are overlain and underlain by relatively less deformed beds. However, the cleavage in the folded units is not as intense as in the enclosing beds, suggesting that the presence of these structures may represent a preservation phenomenon due to the heterogeneity of the deformation related to thrusting rather than slump structures.

The lack of regional folding in the Keppoch block may also be in part attributed to the competence of the rocks of the Keppoch Formation. Turbidites of the James River and South Rights formations which drape the Keppoch block and which are relatively undeformed compared to those same formations in the Georgeville, Maple Ridge and Clydesdale blocks may have been protected from the intensity of the deformation by the underlying Keppoch Formation. Thus the entire Keppoch block may have behaved as a rigid brittle block during deformation. The contrast in style of deformation from north to south in the Antigonish Highlands may also be a reflection of the proximity to a basal thrust during D_1 deformation. The present sharp contact between the areas with these contrasting styles (i.e. the Browns Mountain and Marshy Hope faults) is attributed to later movement on these faults.

Conclusions

The stratigraphy of the Georgeville Group records the progressive development of a marine basin. Geochronological data constrain the depositional age to ca. 618 to 610 Ma. Subaerial to shallow marine volcanic-dominated sequences are overlain by a thick sequence of turbidites and minor mafic volcanic rocks. Sedimentation was synorogenic, derived from flanking volcanic arcs and may have been accompanied by transcurrent faulting. The geochemistry of the volcanic rocks suggests that the basin was genetically associated with a volcanic arc and possibly represents an intra-arc basin. Three distinct magma types may be distinguished: (i) within-plate mafic volcanic rocks that vary from a continental to oceanic setting and have both tholeiitic and alkalic affinities; (ii) basaltic andesites that have a predominantly calc-alkalic affinity; and (iii) felsic volcanic rocks that have volcanic arc compositions. Closure of the basin resulted ca. 615-610 Ma in polyphase deformation involving north-south isoclinal folds and thrusts refolded by upright north-south coaxial folds. A rigorous kinematic model for this deformation is not possible with the available data.

CAMBRO-ORDOVICIAN ROCKS

Stratigraphy

Introduction

Cambro-Ordovician rocks in the Antigonish Highlands are limited to a small area bounded by the Hollow Fault to the northwest and the Greendale Fault to the south, although the latter is locally overstepped and overthrust (Fig. 31). These rocks were previously mapped as part of the Browns Mountain and Arisaig groups (Benson, 1974); however, during the remapping we discovered fossils identified as Early Cambrian (Landing et al., 1980). This has led to a complete revision of the stratigraphy and map boundaries, as reported in this paper. These rocks unconformably overlie the late Precambrian Georgeville Group from which they are lithologically, stratigraphically, structurally and geochemically distinct (Murphy and Keppie, 1987; Murphy et al., 1990). They are overlain with angular unconformity by Late Ordovician to Early Devonian rocks of the Arisaig Group.

The Cambro-Ordovician rocks in the Antigonish Highlands are divided into two groups: the predominantly sedimentary Iron Brook Group and the mainly volcanic McDonalds Brook Group (Fig. 31). The recovery of Early Cambrian fauna from both groups indicates that they are facies equivalents of one another. These groups were subsequently brought together by thrusting (Fig. 31). Iron Brook is the type section of two of the three formations of the Iron Brook Group. The McDonalds Brook Group is named after the brook of that name which runs just south of Sugarloaf Hill (Fig. 32). Type sections for each formation of the McDonalds Brook Group occur in Arbuckle and Malignant Cove brooks (Fig. 32). Gaps in the Cambro-Ordovician fossil record suggest that there is at least one break in the stratigraphic sequence.

Iron Brook Group

Black John Formation

Field relationships. The Black John Formation outcrops mainly in the southwestern end of the area underlain by Cambro-Ordovician rocks (Fig. 31). Its type section occurs in the westward-flowing branches of Doctors and Iron brooks, with other good sections on MacKinnons Brook and in Little Hollow (Fig. 32). The formation consists predominantly of red conglomerates and sandstone, and red and green siltstones and slates with minor rhyolites. The base of the formation is exposed on the western branch of Doctors Brook (Fig. 31 and 33). Here a 6 m thick red rhyolitic tuff overlies green slates of the Maple Ridge Formation, one of the lower units of the late Precambrian Georgeville Group. The green slates dip steeply to the southeast, whereas the rhyolite dips to the north-northwest and overlying conglomerates young northward. Although no significant discordance is evident at the contact, consideration of the regional stratigraphy indicates that this contact represents a major unconformity. The top of the Black John Formation is also exposed in the type section, where it is defined at the base of the lowest limestone bed.

The thickness of the Black John Formation appears to be up to 500 m in the western branch of Doctors Brook. However, a determination of the exact thickness is complicated by the presence of small thrusts which become more common toward the top of the unit, and a displacement of unknown magnitude along the high angle Greendale Fault.

In the western branch of Doctors Brook, the basal rhyolite tuff is an unconformity whereas in the western branch of Iron Brook a thin rhyolitic unit is in fault contact with the Maple Ridge Formation to the south. The rhyolites are massive and aphanitic, but have a brecciated appearance in the field. The rhyolite is concordantly overlain by red conglomerates which are well sorted, polymictic, clast supported, trough cross-bedded, cleaved and contain subrounded to subangular pebbles (up to 1 cm in size) of red siltstone, red slate, metabasalt, green slate, granite, quartz and perthitic feldspar. Some of these pebbles are similar in composition to the underlying Georgeville Group. The conglomerates are interbedded with red cross-bedded sandstones. These conglomerates and sandstones become less common upward as mottled red and green, penetratively cleaved siltstones and slates displaying cross-lamination become more common. The colouration of these rocks is very irregular and is attributed to postdepositional reduction of originally red rocks. Some elliptical (2:1) reduction spots with subhorizontal long axes occur on the cleavage planes. These rocks are in turn overlain by a 3 m thick, massive pale pink to white quartzite displaying load structures along its base. The quartzite is overlain by more red and green slates with fewer siltstones. Just to the east, a 1 m thick felsic tuff is interbedded with these slates on the northern bank of the Little Hollow, but is absent in the type section. Also a gabbroic sill intrudes the succession here close to the contact between the Black John and Little Hollow formations. Above the gabbroic sill the red slates are interbedded with fossiliferous limestone beds. The base of the lowest limestone defines the top of the Black John Formation.

Although the Black John Formation is not fossiliferous, it conformably underlies the fossiliferous late Early Cambrian Little Hollow Formation, and so is probably Eocambrian to Early Cambrian in age.

Petrography. Felsic volcanic rocks occurring at the base of the Black John Formation contain fragments and whole microphenocrysts of albite, sericitized orthoclase and angular grains of quartz. The matrix has been extensively recrystallized and consists of calcite, sericite, albite, quartz, pyrite and minor epidote.

The red conglomerates contain a variety of pebble lithologies, many of which possess an inherited foliation at diverse orientations relative to the foliation in the matrix. The pebbles consist of mafic to intermediate volcanic rocks, granite (which commonly displays graphic textures), siltstone and slate. These pebbles lie in a foliated matrix of quartz, perthitic feldspar clasts, sericite, chlorite and biotite. Most of these pebbles were probably derived from the underlying Georgeville Group, while late Precambrian plutons in the Antigonish or Cobequid highlands or Cape Breton Island may be the source of the perthitic feldspars and graphic granite.

The slates consist of minute grains of quartz and feldspar, commonly with a hematitic coating, surrounded by aligned sericite defining the slaty cleavage.

Environment of deposition. The red colour of the conglomerates, sandstones, siltstones and slates and the lack of interbedded marine rocks suggests that the Black John Formation is terrestrial. The fining upward character of the formation together with its primary structures indicate a fluvial environment, probably in a braided river setting.

Little Hollow Formation

Field relationships. The Little Hollow Formation is best exposed in the type section along the west branch of Iron Brook (Fig. 31 and 32). It also outcrops on the east branch

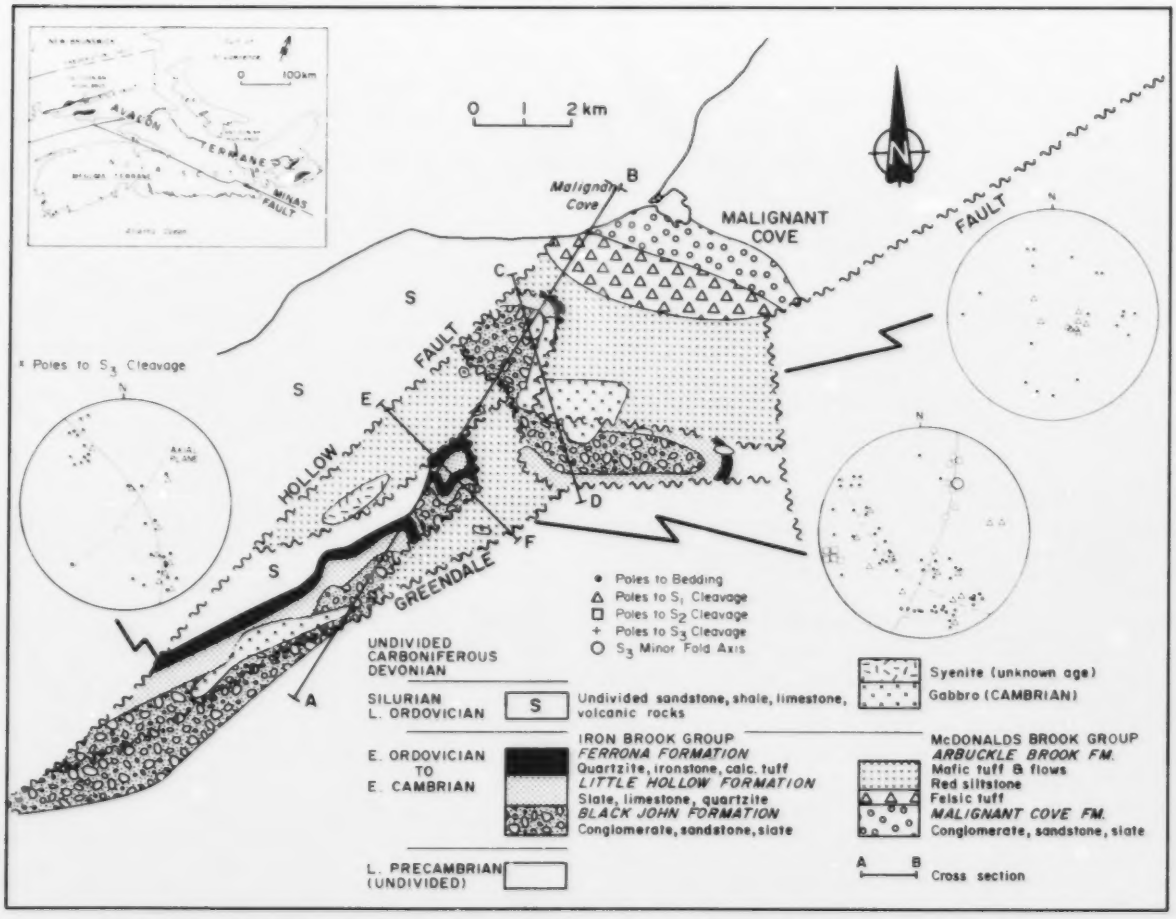


Figure 31a. Simplified geological map of the Cambrian to Lower Ordovician rocks in the Antigonish Highlands (see Map 1749A for details). Stereoplots show structural data from the western, central and eastern parts of the area. The locations of the three cross sections are shown on the map. The inset shows the distribution of Cambro-Ordovician rocks in Nova Scotia.

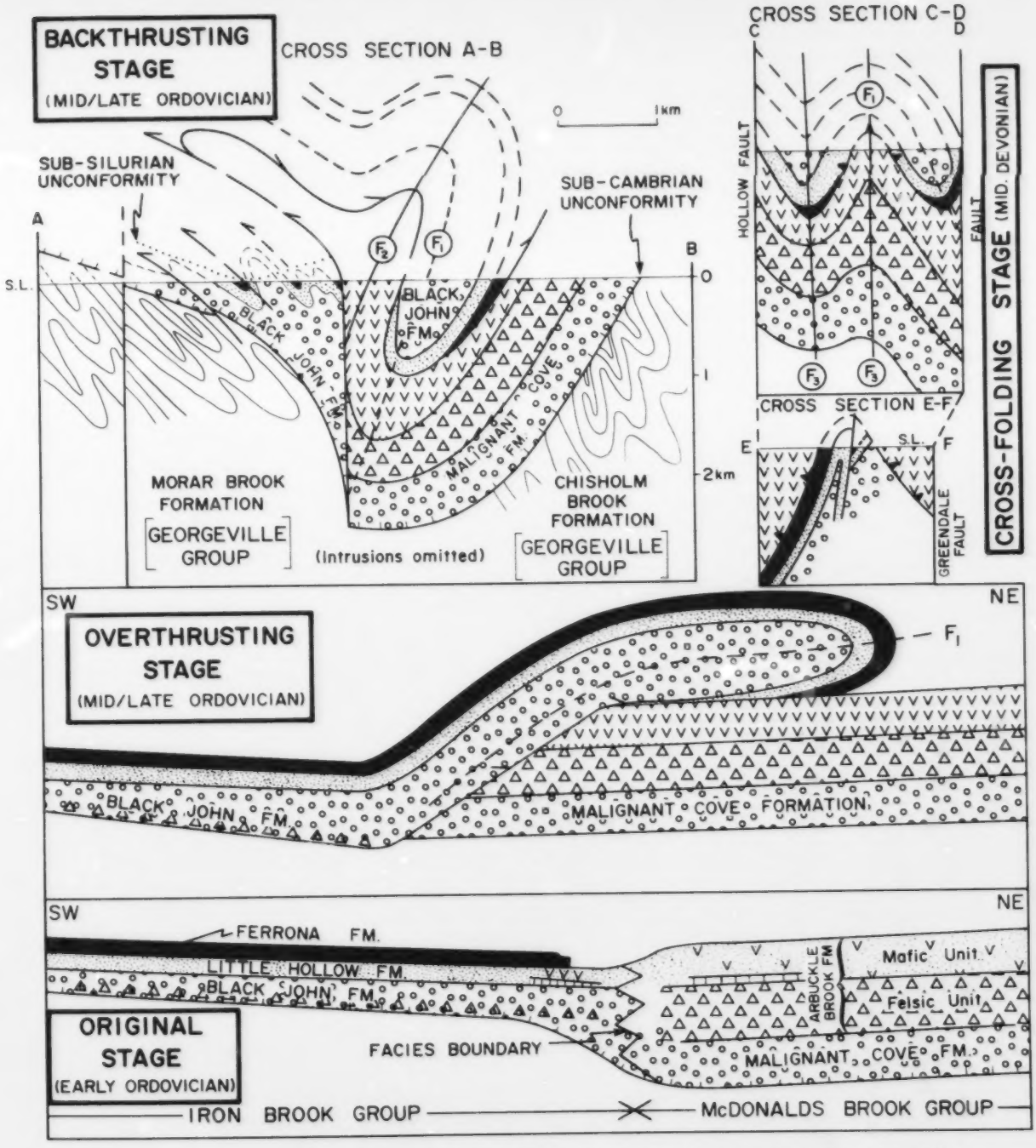


Figure 31b. Successive stages of deformation are shown as: (i) the original stage immediately following deposition of Cambrian-Early Ordovician rocks, depicting the facies change between the Iron Brook and McDonalds Brook groups; (ii) the overthrusting stage (Middle to Late Ordovician) in which the Iron Brook Group has been thrust northeastward over the McDonalds Brook Group; (iii) the backthrusting stage depicted by cross section A-B where thrusting to the southwest is accompanied by folding, verging outward from a central zone; (iv) a cross-folding stage (Middle Devonian) depicted by cross sections C-D and E-F, where northeast-southwest folds deform the rocks. Location of sections shown in Figure 31a.

of Doctors Brook and in isolated exposures around Sugarloaf Hill. The name is derived from the Little Hollow which joins the headwaters of east and west branches of Iron Brook. The term Little Hollow Formation was first used by Benson (1974). It is redefined here and it is applied to a more restricted sequence of rocks. The formation consists predominantly of interbedded red slates and pink limestones, with minor calcareous tuff. The base and top of the formation are here defined by the first and last occurrences of limestone, respectively. The lower contact is exposed in the west branch of Iron Brook and appears concordant and conformable. The upper contact with its overlying Ferrona Formation exposed in both the east and west branches of Iron Brook is concordant.

The thickness of the Little Hollow Formation is impossible to measure due to the numerous thrusts that occur in the type section and elsewhere. In the type section, the Little Hollow Formation has a structural width of 220 m decreasing to as little as 100 m to the northeast.

In the type section, the Little Hollow Formation typically consists of 3 to 5 cm thick interbedded red slates and pale pink fossiliferous limestones (Fig. 33). The red slates are similar to those in the underlying Black John Formation. The limestones are generally fine grained, massive and contain abundant fossil debris. In places, nodules of limestone occur in the red slate matrix. The limestones are preferentially weathered relative to the slates. A relatively thick (2 m) pink limestone occurs near the top of the formation. The calcareous tuffs contain fine grained green and coarse purple and grey angular fragments up to 2 cm in size.

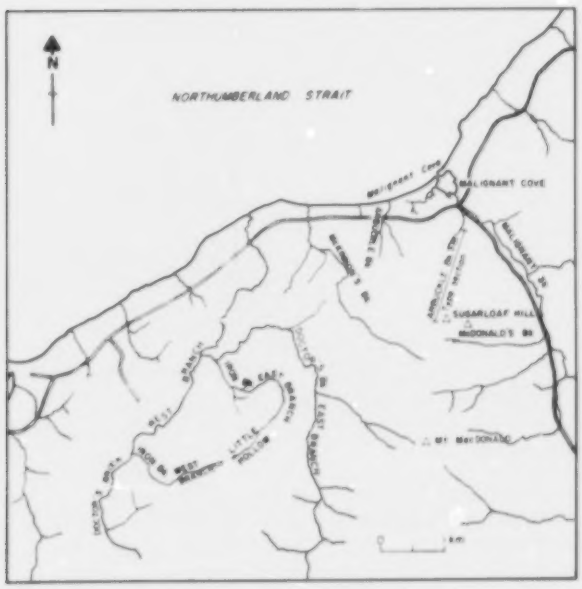


Figure 32. Geography of the Malignant Cove area showing the locations of the place names and type sections mentioned in the text.

Petrography. The petrography of the red slates is similar to those in the Black John Formation. The limestones are generally very fine grained and contain abundant ragged and interlocking grains of calcite, with minor quartz. Many of the grains are coated with hematite. Trilobites, brachiopods and other fossils were observed in thin section (Fig. 34).

The calcareous tuffs contain clasts up to 3 cm in diameter set in a highly altered carbonate matrix. The clasts are predominantly composed of calcite and are often flattened with their long axes usually parallel to a well-defined foliation. The matrix consists of calcite and chlorite generally coated with hematite. The matrix is intensely deformed displaying two fabrics intersecting at an acute angle. Small (<0.5 mm) globules of cryptocrystalline material interpreted as volcanic clasts are also common. The extent of alteration and deformation in this rock is so severe that the identification of the protolith remains problematical.

Age. During 1977, the richly fossiliferous pink limestones were discovered and these samples along with subsequent collections were sent for identification by Landing et al. (1980). The samples yielded the following assemblage of trilobites, brachiopods and problematica. Source references for these fossils are given in Landing et al. (1980).

Trilobites

Acanthomicmacca sp. aff. spinosa (Cobbold, 1910)
A. walcotti (Matthew, 1899)
?Avalonia sp.
Callavia broeggeri (Walcott, 1888) (=*callavei* Lapworth, 1888)
Comleulla sp.
Ladadiscus sp. indet.
Serrodiscus bellimarginatus (Shaler and Foerste, 1910)
Strenuella strenua (Billings, 1894)
Triangulaspis vigilans (Matthew, 1899) (=*annia* Cobbold, 1910)

Brachiopods

Lingulella sp. s.l.
Mickwitzia sp.

Problematica

Amphigeisina? sp.
Eccentrotheca kanesia n. gen., n. sp.
Lapworthella sp.
Pelagiella sp.
Rhombocorniculum cancellantum (Cobbold, 1921)
Sunnaginia imbricata (Missarzhevski, 1969)
Turcutheca? sp. *operculum*
Problematicum 1

The trilobites may be referred to the Callavia Zone (s.l.) typical of the Avalon Zone of late Early Cambrian, and contain both cosmopolitan and provincial facies-controlled elements (Landing et al., 1980).

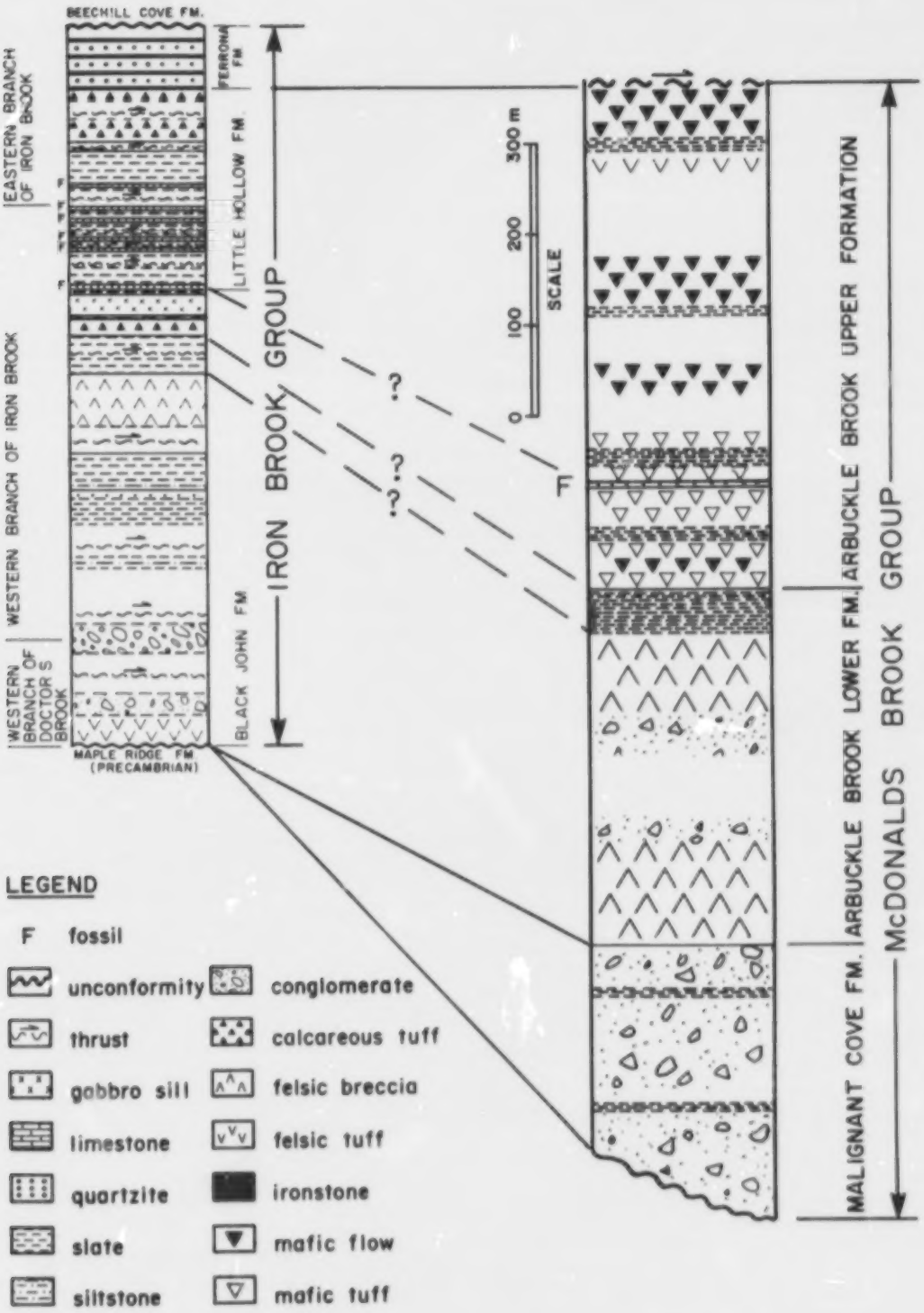


Figure 33. Stratigraphic columns of the Iron Brook and McDonalds Brook groups.

Environment of deposition. The alternating red slates and fossiliferous limestones and the rapid facies changes indicate a nearshore shallow marine environment in which seasonal fluctuations produced influxes of red terrestrial clay and quiescent clear water periods. That this shelf was connected to the larger oceans is indicated by the presence of cosmopolitan open-ocean tribolites.

Ferrona Formation

Field relationships. The Ferrona Formation is best exposed in the type section, which is located in the east branch of Iron Brook (Fig. 31 and 32). It also outcrops along the eastern bank of East Doctors Brook and in the many pits and trenches excavated to mine the ironstones typical of the unit. The base of the formation is defined at the top of the highest interbedded limestone and slate of the underlying Little Hollow Formation. The contact is exposed in the eastern branch of Iron Brook and is concordant. The top of the Ferrona Formation is unconformably overlain by the Early Silurian Beechill Cove Formation in the eastern branch of Iron Brook.

The presence of thrusts in the unit precludes any estimate of the thickness of the formation. Its structural width ranges from about 50 to 70 m. In the type section, the Ferrona Formation consists of interbedded massive grey quartzites and at least four 1 to 3 m thick oolitic ironstones (Fig. 33). The quartzites are generally devoid of primary structures and their contacts with the ironstones may be either sharp or gradational. The ironstones vary from relatively pure and oolitic-pisolitic, to containing much well-sorted clastic material up to 5 mm in size consisting of quartz, feldspar, red slate and jasper. Inarticulate brachiopod and trilobite fragments are usually also present in the ironstones.

Figure 34. Photomicrograph (4 mm wide) of the limestones of the Little Hollow Formation showing a cross section through a trilobite. For further details see Landing et al. (1980).

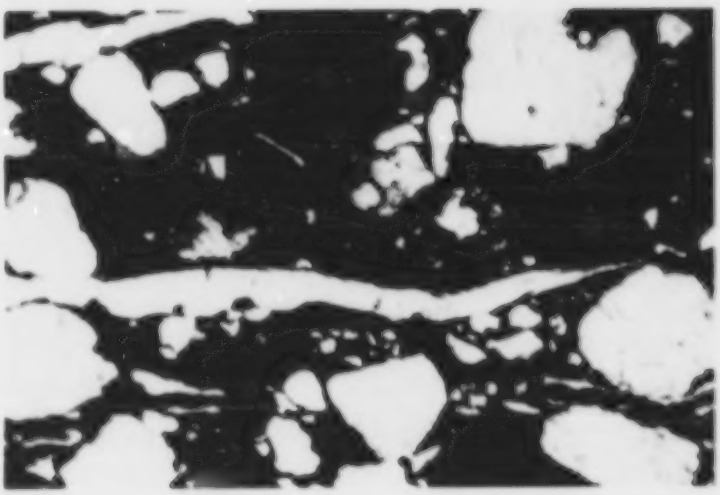


Figure 35. Photomicrograph (4 mm wide) of the ironstones of the Ferrona Formation. Debris of brachiopods (elongate fragment in the centre of the photograph) and clasts of quartz with hematite-filled cracks (e.g. upper right) are enclosed in a hematitic matrix.

Petrography. The quartzites consist of well-sorted, less than 3 mm, subangular to subrounded undulose quartz grains displaying quartzose overgrowths set in a quartz matrix. Minor hematitic alteration is present.

The impure ironstones consist of quartz, jasper, inarticulate brachiopod and red slate fragments set in a hematite matrix (Fig. 35). The subangular to subrounded quartz grains display many optically continuous overgrowths separated by thin hematitic coatings. The quartz grains often display undulose extinction and contain innumerable minute cracks filled with hematite. Thin quartz needles may also be observed extending into the hematitic matrix. This feature is interpreted as an indication of instability between quartz and hematite.

The oolitic-pisolitic ironstones are texturally distinguished from other ironstones by the relative lack of clasts. They are dominated by ooids (<5 mm) of hematite commonly surrounding a clastic fragment at the core.

Age. The Ferrona Formation concordantly overlies the Lower Cambrian Little Hollow Formation and is unconformably overlain by the Lower Silurian (Llandoverian) Beechill Cove Formation. The ironstones contain much fragmentary inarticulate brachiopods identified as *Obolus (lingulobus) spissa* and *Lingulella* (?) (Williams, 1914). Despite extensive additional collection no specimens suitable for identification were found (G.S. Nowlan, pers. comm., 1984). These inarticulate brachiopods are long ranging. However, the close lithostratigraphic correlation between the Cambro-Ordovician rocks of the Antigonish Highlands and southeastern Newfoundland suggests that the Ferrona Formation is equivalent to the Wabana Group (Hayes, 1915; Ranger, 1979). The Arenigian age of the Wabana Group is defined by graptolites which suggest a similar age for the Ferrona Formation. It follows from this age assignment that there is a significant gap in the stratigraphic record between the Early Ordovician Ferrona and Early Cambrian Little Hollow formations. This suggests the presence of a disconformity or a thrust within the Iron Brook Group with the Middle and Upper Cambrian absent.

Environment of deposition. The close association of orthoquartzites and fossiliferous oolitic-pisolitic ironstones suggests a marginal marine environment. The quartzites may represent beach or sand bank deposits, whereas the oolites suggest deposition in water constantly agitated by tidal currents. The pisolites may indicate the presence of blue-green algae (Dimroth and Kimberly, 1976). The oolites and pisolites may have been washed into sand bars by storm or tidal currents. An oxidizing environment is indicated by the predominance of hematite. The source of the iron is uncertain, but the ironstones contain high Ti and V, suggesting that the hematite was originally deposited as magnetite from a volcanogenic source (e.g. Miyashiro and Shido, 1975). Alternatively, the iron may have been leached from supratidal marshes surrounding the marine basin (cf. Ranger, 1979).

McDonalds Brook Group

Malignant Cove Formation

Field relationships. The Malignant Cove Formation was first named by Williams (1914) for the red conglomerates around Malignant Cove (Fig. 31 and 32). It is best exposed in Malignant Brook and its tributaries, which are here taken as the type section. In a tributary of Malignant Brook, a gap in the exposure of several metres separates the lowest outcrop of the Malignant Cove Formation from the underlying Precambrian Chisholm Brook Formation. It is inferred to be an unconformity because it rests on one of the lowest units of the Georgeville Group and the Malignant Cove Formation contains deformed pebbles of Georgeville Group lithologies. The top of the Malignant Cove Formation is not exposed. The contact with the overlying Arbuckle Brook Formation is defined by the first occurrence of volcanic rocks in the section (Fig. 33). These units appear to be concordant on a regional scale (Fig. 31). Interlayering of conglomerates and volcanic rocks occurs in the vicinity of the contact.

The Malignant Cove Formation is approximately 300 m thick. It is made up predominantly of red conglomerates with minor breccias, sandstones and slates. The clasts in the conglomerates and breccias range from pebble to cobble in size, are typically subangular and are set in a sandy matrix. The formation varies from massive, poorly bedded, imbricate clast-supported conglomerates through planar cross-bedded imbricate sand matrix-supported conglomerates to rare planar and cross-bedded sandstones and fine slates. The formation as a whole does not show any upward fining or coarsening. On an outcrop scale, the lithology shows both upward fining and coarsening.

Petrography. The conglomerates contain a wide range of clasts. Mafic and felsic volcanic clasts are predominant and include altered basalt, porphyry, rhyolite and different kinds of tuff. Clast lithologies also include green sericitic slate, red siltstone, wacke, granite, chert, jasper, perthitic feldspar and albite. Many of these clasts may be derived from the underlying Precambrian Georgeville Group, or the neighbouring Cobequid Highlands and Cape Breton Island. These clasts are set in a matrix composed mainly of quartz with minor sericite, chlorite and biotite. Cleavage in the pelitic clasts has diverse orientations relative to the fabric in the matrix. This suggests that the cleavage was present in the clast lithologies prior to being incorporated in the conglomerates.

Age. The Malignant Cove Formation was inferred to be Middle to Late Ordovician in age by Boucot et al. (1974) and Benson (1974). Since it is not fossiliferous, its age must be deduced from its relationships with other units in the area. It postdates the Precambrian Georgeville Group and some of the deformation experienced by these rocks because it contains deformed Georgeville clasts. On the other hand, it concordantly underlies the Arbuckle Brook Formation from which late Early Cambrian fossils have been recovered. Thus, the age of the Malignant Cove Formation is probably Eocambrian to Early Cambrian.

Environment of deposition. The red colour of the Malignant Cove Formation and the lack of interbedded marine strata suggest a terrestrial environment of deposition. The nature of the conglomerates indicates that they are fluviatile deposits probably laid down in a proximal braidplain environment. The lack of any overall fining or coarsening upward suggests that the source area was periodically rejuvenated possibly by uplift along faults during the formation of the basin.

Arbuckle Brook Formation

Field relationships. This unit is best exposed on Sugarloaf Hill, in a tributary of Malignant Cove Brook and in the East Doctors Brook area (Fig. 31 and 32); however, these names have been used previously for other formations. This leaves Arbuckle Brook, in which the unit is relatively poorly exposed, as the namesake. The lack of any continuous section across this unit means that the type section is composite. It is made up from sections exposed in a northerly flowing tributary of Malignant Brook and an overland traverse up Sugarloaf Hill (Fig. 32). The Arbuckle Brook Formation is divided into two members: a lower member consisting of felsic volcanic rocks and interlayered minor red conglomerates, and an upper member characterized by mafic volcanic rocks interbedded with relatively subordinate slates and pink limestones (Fig. 36). The base of the formation is placed at the first appearance of felsic volcanic rocks. The contact with the Malignant Cove Formation is not exposed but appears to be concordant and transitional on a regional scale (Fig. 31). The boundary between the lower and upper members is placed at the base of the lowest mafic volcanic rock (Fig. 33). It is not exposed but coincides with a magnetic gradient (Geological Survey of Canada, 1982). The top of the Arbuckle Brook Formation is truncated by thrust faults (Fig. 31). Magnetic gradients within the upper member of the Arbuckle Brook Formation trend approximately parallel to the underlying unit boundaries suggesting a reasonably continuous sequence. The thickness of the Arbuckle Brook Formation is estimated to be about 350 m thick.

The lowermost units consist of massive subaerial felsic lithic tuffs with a few interbedded rhyolitic flows and red conglomerates. A gap in exposure separates these rocks from more felsic tuffs near the top of the lower member. The tuffs are typically pale green and contain purple to green angular lithic fragments up to 4 cm in size, and plagioclase phenocrysts.

About 1 km north of Sugarloaf Hill felsic volcanic rocks of the lower member are overlain by the upper member (after a gap in exposure) consisting of green aphanitic submarine mafic alkalic within-plate flows (Murphy et al., 1985) grading laterally into hyaloclastites and minor tuffs containing elliptical porphyritic lithic fragments (<2 cm), red slate and pink fossiliferous limestone of the upper member. Intermittent outcrops of mafic volcanic rocks occur to the south. The mafic volcanic rocks are well exposed around the top of Sugarloaf Hill. Here, a mafic tuff unit is interbedded with relatively minor red slates and siltstones, pink fossiliferous limestones and mafic flows. This area coincides with an area of high magnetic gradient interpreted to be due to a greater proportion of mafic rocks. Mafic dykes and gabbro stocks are spatially associated with the mafic upper member of the Arbuckle Brook Formation, suggesting that they are the subvolcanic feeders for the extrusive rocks. Similar mafic volcanic rocks interbedded with minor red slates also outcrop in and adjacent to East Doctors Brook (Fig. 31 and 33).

Petrography. The sedimentary rocks in the Arbuckle Brook Formation are very similar to lithologies already described from other Cambrian units. Thus, the red conglomerates are the same as those in the Malignant Cove Formation, while the red slates and pink fossiliferous limestones are identical to those in the Little Hollow Formation. These descriptions will not be repeated here.

The rhyolites consist of microphenocrysts of saussuritized plagioclase (now albite) and minor strongly sericitized orthoclase in a fine grained matrix of albite, potassium feldspar, quartz, sericite and chlorite with minor epidote and calcite. Thin rod-like microlites of feldspar display a good flow alignment. Quartz is typically interstitial and its shape is largely controlled by that of the feldspar. Sericite and chlorite define a nonpenetrative foliation in these rocks. The tuffs consist of green to purple lithic fragments up to 3 cm in diameter enclosed by a fine grained highly altered matrix. These fragments contain euhedral albite up to 6 mm in length, palagonite, minor chlorite and epidote. The matrix is exceedingly fine grained and is composed of quartz, albite and sericite with minor calcite and chlorite.

In the mafic rocks, volcanic textures appear to have been well preserved, but subsequent metamorphic events have affected the original mineralogy. The flows are generally fine grained and contain slender microlites of partially sericitized albite, which exhibits a pilotaxitic texture, set in a matrix of quartz, epidote, calcite, chlorite, sphene, titanomagnetite and sericite ± rutile. Local augite

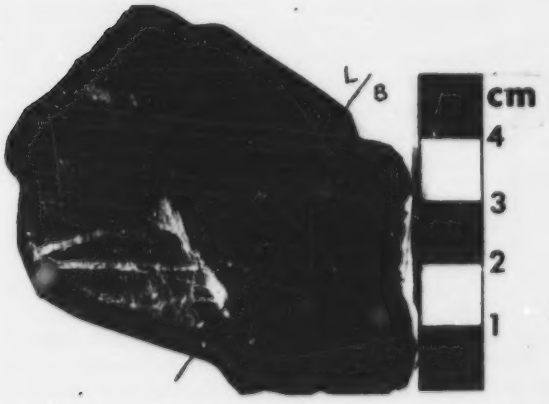


Figure 36. Interlayered basalt and limestone, Arbuckle Brook Formation. The limestone (left) is identical to that of the Little Hollow Formation. The mafic volcanic (right) is a hyaloclastite. Clasts of mafic volcanic rock in a fine grained mafic matrix are visible. L-limestone, B-basalt. GSC 1991-185.

phenocrysts generally altered to chlorite may be present. The secondary mineralogy reflects a greenschist facies metamorphism.

The hyaloclastites contain mafic lithic fragments enclosed in a fine grained, sometimes intensely altered matrix. The fragments range from coarse ash (1 mm) to lapilli (20 mm) in length and vary in shape from subrounded or subelliptical to highly irregular. The mineralogy of the fragments and of the matrix is essentially similar to that of the mafic flows, but they are finer grained. In most fragments, albite displays a preferred flow orientation, parallel to the long axis of the fragment. This feature probably indicates a primary flattening, imparting a weak nonpenetrative foliation. More ragged fragments tend to contain vesicles that are filled with chlorite. Hyaloclastites are distinguished from tuffs by the presence of flow textures in their matrix. However, in some samples, the matrix was intensely altered, and the distinction could not be made.

Age. The age of the Arbuckle Brook Formation is given by the late Early Cambrian age of the fossils in the interbedded thin pink limestone(s) identical to those in the Little Hollow Formation.

Environment of deposition. The geochemistry of the volcanic rocks of the Arbuckle Brook Formation suggests that they were extruded in a within-plate environment undergoing a minor localized rifting event (Murphy et al., 1985). This is consistent with the subaerial to shallow marine transition observed in the interbedded sediments of the Arbuckle Brook Formation.

Correlations

The recovery of Early Cambrian fauna from both the predominantly sedimentary Iron Brook Group and the mainly volcanic McDonalds Brook Group indicates that they are correlatives which have been juxtaposed by thrusting (Fig. 31 and 33). The fossiliferous pink limestones are the best marker horizons for correlative purposes; however, the volcanic units may also be used. Any detailed correlation of these two groups is impossible because it is likely that the volcanic edifice forming the McDonalds Brook Group would not be submerged at the same time as surrounding nonvolcanic areas. The felsic volcanic rocks in the Black John Formation may possibly be correlated with those in the lower member of the Arbuckle Brook Formation (Fig. 33). These tentative correlations suggest that: (i) the Black John Formation is equivalent to the lower part of the Arbuckle Brook Formation and part of the Malignant Cove Formation; and (ii) the Little Hollow Formation is probably equivalent to most of the upper member of the Arbuckle Brook Formation (Fig. 33).

These correlations suggest a major facies change from a predominantly volcanic environment in the northeast to a predominantly sedimentary environment in the southwest. This suggests that the locus of volcanism was located closer to the northeastern end of the basin. It is postulated that the volcanism was induced by dextral movements on northeast-southwest-trending faults which produced a pull-apart basin where the Greendale Fault curves to meet the Hollow Fault (Fig. 31). The facies distribution suggest that northwest-southeast fissures parallel to the rift axis were located at the northeastern end of the basin. These fissures are inferred to be a response to stretching of the lithosphere beneath the pull-apart basin.

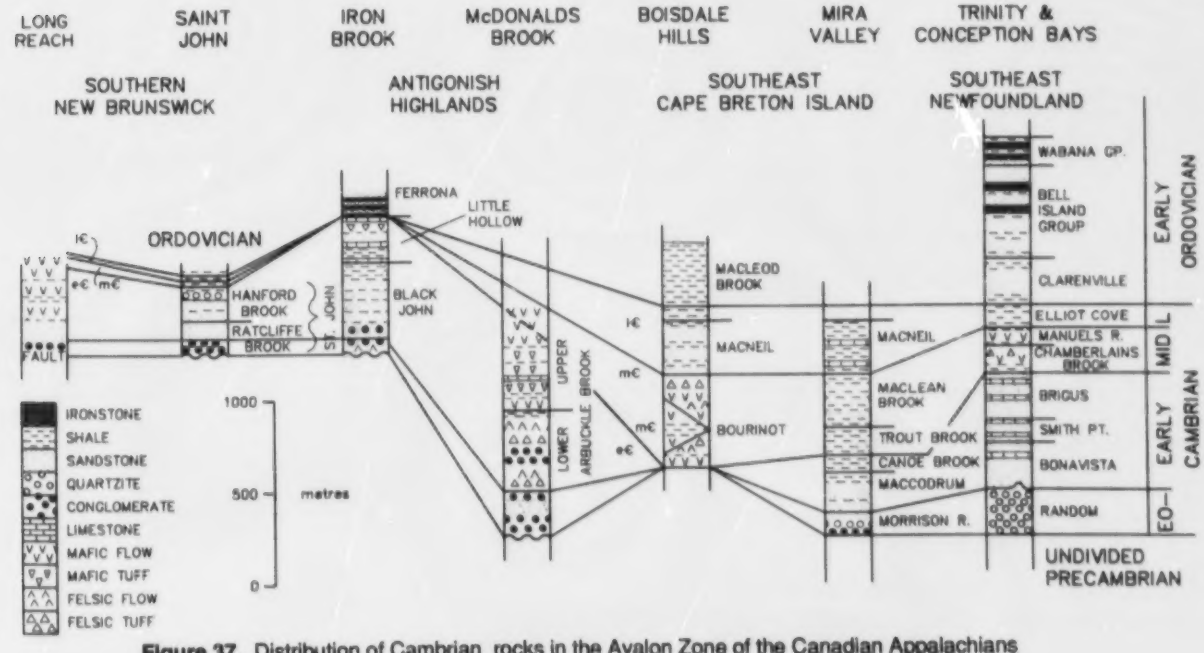


Figure 37. Distribution of Cambrian rocks in the Avalon Zone of the Canadian Appalachians (from Williams, 1978).

Table 4. Geochemistry of the mafic volcanic rocks of the Arbuckle Brook Formation

	E16-343	E16-309	E16-308	E16-448	E16-313	E16-517	E16-342	E16-426	E16-424	E16-475	E16-376	E16-301
SiO_2	35.9	41.7	42.8	42.8	43.1	43.3	43.8	43.8	43.8	44.0	44.1	44.5
TiO_2	0.92	3.76	3.38	3.75	2.26	2.49	3.49	3.67	3.74	3.53	3.56	2.49
Al_2O_3	7.31	15.5	14.8	15.9	15.0	12.9	14.8	14.8	15.2	15.1	15.6	14.6
Fe_2O_3	1.79	5.50	6.72	10.06	2.28	3.63	5.28	2.11	5.17	10.50	5.83	5.67
FeO	5.05	6.25	6.85	2.35	8.85	3.60	8.65	10.20	8.55	3.75	7.05	9.90
MnO	0.37	0.62	0.31	0.65	0.48	0.23	0.28	0.34	0.53	0.28	0.51	0.43
MgO	5.19	8.13	6.36	6.75	5.88	1.59	4.36	5.25	4.77	6.40	6.11	6.64
CaO	22.7	6.18	9.07	7.04	7.79	14.8	6.33	5.07	5.30	8.06	4.86	4.68
Na_2O	0.80	3.48	3.28	3.69	3.74	6.02	4.15	3.06	4.21	2.77	4.23	3.67
K_2O	0.06	0.59	0.45	0.75	0.50	0.25	2.07	2.74	1.08	0.88	0.35	0.51
P_2O_5	0.11	1.01	0.92	0.82	0.48	0.59	0.69	0.76	0.76	0.94	1.07	0.32
H_2O^+	3.10	4.55	3.10	3.80	4.05	1.65	4.75	4.30	4.90	2.90	4.10	4.80
H_2O^-	0.51	0.00	0.65	0.34	0.89	2.92	0.00	0.09	0.32	0.14	0.03	0.00
CO_2	16.1	0.05	0.60	1.50	3.30	7.05	1.45	<0.05	0.15	0.10	0.05	2.90
TOTAL	98.92	97.32	99.20	100.20	98.60	101.02	100.1	96.21	98.48	99.33	97.45	100.91
Ni	n.d.	106	106	n.d.	128	11	31	21	29	120	74	129
Cr	n.d.	111	145	n.d.	302	91	47	22	86	236	143	257
Rb	6	36	28	45	21	12	38	40	36	48	28	24
Sr	189	803	849	867	256	219	576	654	876	946	673	375
Nb	4	33	33	30	16	14	19	21	23	34	33	12
Y	23	57	54	54	30	28	47	55	41	45	49	22
Zr	75	356	311	322	160	220	254	273	291	317	360	132
La	8	39	43	42	19	18	26	38	27	46	38	17
Ce	13	88	99	94	42	45	68	82	69	105	89	33
Nd	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Sm	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Eu	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Tb	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d
Yb	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d
Lu	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Sc	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.

	E16-427	E16-303	E16-476	E16-375	E16-801	E16-802	E16-803	E16-804	E16-805	E16-806	E16-807
SiO_2	44.9	45.3	45.8	45.8	48.00	47.55	44.40	47.51	45.51	45.03	44.51
TiO_2	3.43	1.63	3.53	3.03	2.47	2.39	1.50	1.32	2.56	2.89	3.64
Al_2O_3	15.0	15.4	15.7	15.2	16.42	15.49	16.72	14.40	16.69	16.00	17.40
Fe_2O_3	4.39	5.11	5.84	6.61	4.89	2.92	3.47	1.20	6.23	5.16	5.40
FeO	9.45	12.6	4.15	5.65	6.44	6.01	8.16	8.95	5.85	5.92	5.55
MnO	0.31	0.47	0.63	0.53	0.26	0.21	0.19	0.17	0.37	0.25	0.22
MgO	4.93	7.67	6.94	6.02	5.06	4.41	5.61	5.24	7.47	6.60	6.32
CaO	5.49	2.13	8.09	6.43	6.26	9.14	5.53	7.17	7.81	8.62	9.29
Na_2O	3.47	3.43	3.18	4.30	3.19	4.15	3.14	5.18	2.81	3.22	3.37
K_2O	2.81	0.10	0.87	0.21	0.94	0.45	0.75	0.54	0.67	0.83	0.72
P_2O_5	0.75	0.31	0.97	1.11	0.57	0.57	0.26	0.25	0.82	0.83	0.84
H_2O^+	3.75	5.90	3.35	3.65	3.27	3.84	3.88	4.28	3.33	3.52	3.11
H_2O^-	0.00	0.00	0.27	2.60	0.07	0.22	0.06	0.01	0.37	0.42	0.20
CO_2	0.85	0.05	0.10	<0.05	1.62	1.12	5.12	4.06	0.08	0.23	0.36
TOTAL	99.53	100.1	99.22	101.14	99.46	98.47	98.79	100.28	100.57	99.52	100.93
Ni	17	85	90	51	38	37	42	39	84	69	71
Cr	42	238	197	102	128	130	163	161	101	109	96
Rb	27	9	48	25	15	14	9	1	9	9	9
Sr	696	163	922	793	43	445	300	348	1085	1056	1065
Nb	19	5	36	38	12	13	7	5	66	71	66
Y	46	22	53	44	40	41	34	28	31	32	31
Zr	253	114	334	375	215	223	103	101	365	356	335
La	29	9	46.5	49.7	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Ce	89	20	106	113	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Nd	n.d.	n.d.	55.4	58.0	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Sm	n.d.	n.d.	11.4	11.2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Eu	n.d.	n.d.	3.47	3.44	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Tb	n.d.	n.d.	1.34	1.35	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Yb	n.d.	n.d.	2.45	2.42	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Lu	n.d.	n.d.	0.37	0.39	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Sc	n.d.	n.d.	23.3	16.3	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.

n.d. = not determined

The Cambro-Ordovician rocks of the Antigonish Highlands are similar in lithostratigraphy and fauna to Cambro-Ordovician rocks described in other parts of the Avalon Composite Terrane (Fig. 37), such as in Cape Breton Island (Hutchinson, 1952), the St. John area of southern New Brunswick (Currie, 1986) and the Inlet Group in the Burin Peninsula of Newfoundland (Strong et al., 1978).

Geochemistry

Analytical notes

A total of 23 samples of mafic volcanic rocks and 12 felsic volcanic rocks were analyzed. These analyses are displayed in Tables 4 and 5 respectively. The major elements were determined either by atomic absorption at the Technical University of Nova Scotia, Halifax, or by X-ray fluorescence and supplementary techniques at X-Ray Assay Laboratories Limited, Toronto. The trace elements Zr, Nb, Y, Ba, Sr and Rb were analyzed by X-ray fluorescence, whereas rare earth elements (REE), Sc, Th and Hf were determined by instrumental neutron activation at St. Mary's University, Halifax. The precision and accuracy of trace element analyses are generally better than 10 percent (Cameron, 1980).

Effects of alteration

The Cambrian volcanic rocks were metamorphosed to the greenschist facies, an event that may have resulted in significant changes in the primary chemistry of the rocks. The extent of alteration is reflected in the range of concentrations of some of the major elements. For example, in the mafic rocks, Na_2O varies from 0.2 to 7.5 weight percent, CaO and K_2O display unusually wide ranges in concentration and FeO/Fe_2O_3 ratios range from 5 to 0.3. The effects of alteration on the geochemistry of these rocks have been evaluated in detail by Murphy and Hynes (1986). The limited ranges in concentrations of the relatively immobile elements in the CO_2-poor rocks, such as Ti, Zr, Y and P, and their consistent interelement ratios, suggest that their relative mobility was negligible.

Mafic rocks

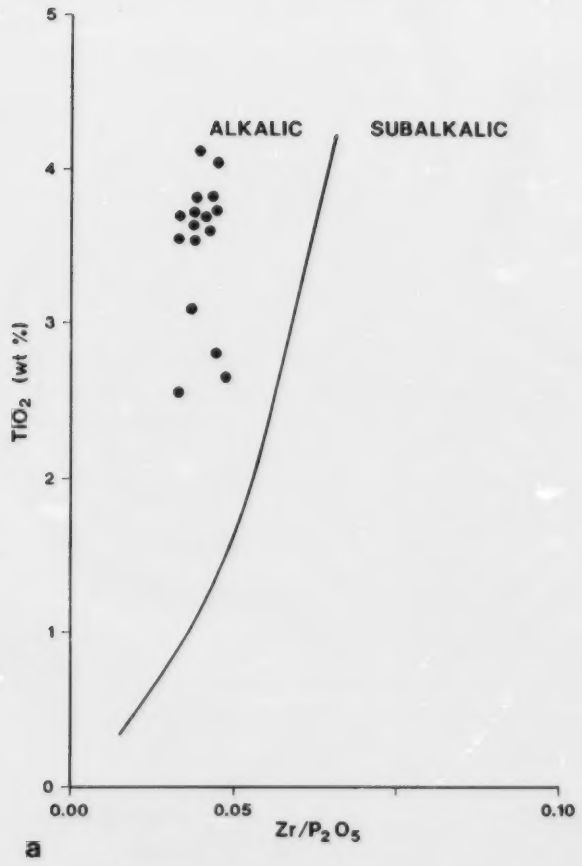
The volcanic rocks are bimodal with respect to SiO_2, being composed mainly of basalts and rhyolites. The basalts have SiO_2 contents (on a volatile-free basis) between 45.0 and 48.3 percent and a FeO^t/MgO ratio between 1.36 and 3.07. They have high contents of TiO_2 ($X \pm 1 = 3.5 \pm 0.69\%$) that increase

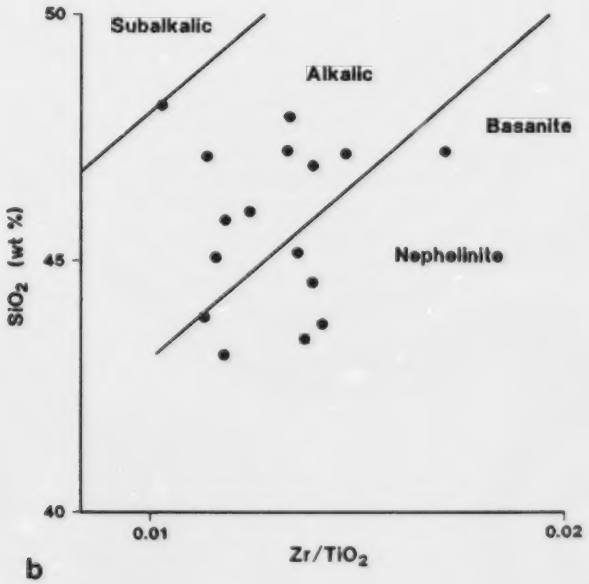
Table 5. Geochemistry of the felsic volcanic rocks of the Arbuckle Brook Formation

	E16-306	E16-338	E16-468	E16-305	E16-339	E16-425	E16-353	E16-354	E16-516	E16-304	E16-307	E16-472
SiO_2	58.1	58.1	58.2	59.7	59.8	61.6	63.7	63.7	69.1	69.5	72.8	73.1
TiO_2	0.96	0.42	0.43	0.56	0.41	0.55	0.64	0.62	0.58	0.32	0.33	0.72
Al_2O_3	16.1	18.6	18.6	12.5	17.8	17.5	15.9	14.1	15.4	14.7	13.9	13.4
Fe_2O_3	2.90	5.09	3.90	1.29	5.72	8.68	3.42	2.62	2.48	2.68	1.34	2.04
FeO	2.90	3.20	3.30	4.80	1.05	0.20	2.5	2.50	0.50	1.80	1.30	0.70
MgO	0.42	0.34	0.22	0.39	0.37	0.35	0.21	0.45	0.27	0.20	0.34	0.28
MnO	1.46	0.70	0.73	2.54	0.36	0.77	1.93	1.58	0.56	0.91	0.83	0.54
CaO	3.94	0.85	0.82	6.80	1.78	0.37	1.55	4.22	0.38	1.83	1.62	0.87
Na_2O	1.48	6.37	6.56	0.85	7.45	3.87	5.67	3.82	2.43	1.97	1.59	4.75
K_2O	2.77	3.62	3.69	1.75	2.07	3.82	1.16	1.89	5.40	1.97	2.25	1.95
P_2O_5	0.35	0.14	0.14	0.20	0.13	0.17	0.14	0.15	0.13	0.09	0.07	0.09
H_{2O}	3.15	1.55	1.50	3.20	0.95	2.00	1.95	1.95	1.45	2.55	2.15	1.00
H_2O-	4.06	0.00	0.00	1.49	0.19	0.08	0.31	0.45	0.21	0.55	0.82	0.25
CO_2	2.55	0.55	0.35	4.25	1.10	0.50	0.65	2.50	0.05	1.05	0.55	0.35
TOTAL	101.14	99.53	98.44	100.32	99.18	100.46	99.78	100.64	98.94	100.12	99.98	100.04
Rb	72	86	85	47	60	106	46	54	204	53	56	72
Sr	99	313	323	142	176	112	188	139	131	128	108	163
Nb	12	63	72	11	60	61	9	8	11	10	10	12
Y	31	54	55	33	43	35	26	20	27	25	24	30
Zr	245	636	638	194	596	578	175	165	233	207	202	244
Ba	408	n.d.	n.d.	262	n.d.	404	268	769	1053	688	372	368
Ni	9	n.d.	n.d.	8	n.d.	7	23	17	5	<5	7	17
Cr	18	n.d.	n.d.	21	n.d.	23	60	56	24	25	30	59
V	63	n.d.	n.d.	18	n.d.	11	100	99	28	10	13	68
Ga	19	n.d.	n.d.	15	n.d.	24	17	15	16	14	14	13
Th	<10	n.d.	n.d.	<10	n.d.	<10	<10	<10	<10	<10	<10	<10
Sc	n.d.	n.d.	n.d.	n.d.	n.d.	4.5	n.d.	n.d.	8.3	n.d.	3.2	n.d.
La	n.d.	n.d.	n.d.	n.d.	n.d.	64.5	n.d.	n.d.	23.2	n.d.	27.8	n.d.
Ce	n.d.	n.d.	n.d.	n.d.	n.d.	143	n.d.	n.d.	53.4	n.d.	63	n.d.
Nd	n.d.	n.d.	n.d.	n.d.	n.d.	56.5	n.d.	n.d.	23.5	n.d.	28.0	n.d.
Sm	n.d.	n.d.	n.d.	n.d.	n.d.	8.77	n.d.	n.d.	4.72	n.d.	5.17	n.d.
Eu	n.d.	n.d.	n.d.	n.d.	n.d.	2.31	n.d.	n.d.	1.32	n.d.	1.28	n.d.
Tb	n.d.	n.d.	n.d.	n.d.	n.d.	1.32	n.d.	n.d.	0.91	n.d.	0.85	n.d.
Yb	n.d.	n.d.	n.d.	n.d.	n.d.	4.59	n.d.	n.d.	3.86	n.d.	3.48	n.d.
Lu	n.d.	n.d.	n.d.	n.d.	n.d.	0.73	n.d.	n.d.	0.66	n.d.	0.59	n.d.

n.d. = not determined

only slightly with increasing differentiation. The alkalic affinities of the mafic flows and feeder dykes are shown on TiO_2 versus Zr/P_2O_5, SiO_2 versus Zr/TiO_2 and P_2O_5 versus Zr (Fig. 38a,b,c) and their chondrite-normalized REE patterns, which display distinct light REE (LREE) enrichment and fractionated heavy REE (HREE) with La/Yb=20 (Fig. 39a). A within-plate tectonic environment is indicated on TiO_2 versus Zr and Zr/Y versus Zr discrimination diagrams (Fig. 40a,b). The range and positive correlation between Ni and Cr in the mafic rocks (Table 4) and the occurrence of olivine, chromite and augite as the cumulus phases in the spatially associated gabbroic stocks and in the mafic dykes suggest that olivine, chromite and augite were fractionating phases. The role of plagioclase is difficult to assess because of the mobility of Ca and Sr. However, the presence of plagioclase phenocrysts in the mafic dykes suggests that plagioclase may also have been a fractionating phase.





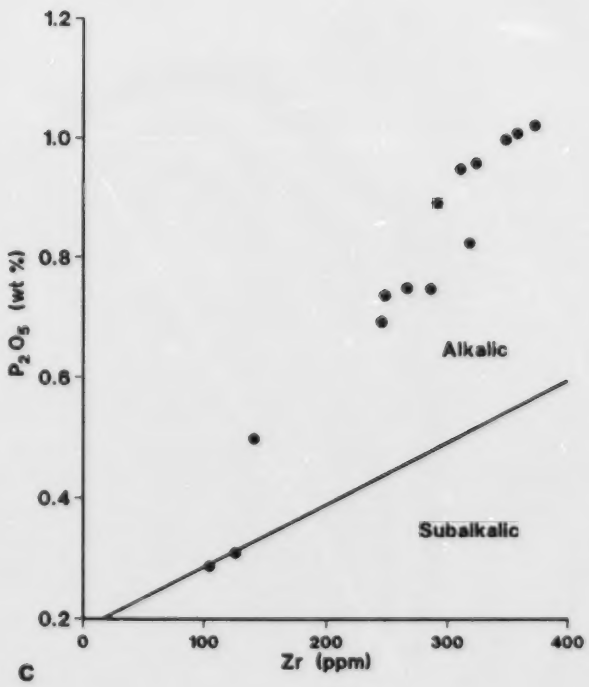


Figure 38. Discrimination diagrams for the mafic volcanic rocks of the Arbuckle Brook Formation, after Floyd and Winchester (1975) and Winchester and Floyd (1977). (**a**) TiO_2 versus Zr/P_2O_5, (**b**) SiO_2 versus Zr/TiO_2 and (**c**) P_2O_5 versus Zr.

Felsic rocks

Most rhyolites, on a volatile-free basis, contain greater than 70 percent SiO_2, greater than 11 percent Al_2O_3 and greater than 80 percent normative quartz + feldspar. They generally have a peraluminous character and contain normative corundum. When plotted on an Ab-Or-Q diagram (Fig. 41) their normative compositions show a large scatter, with many rocks falling outside the granitic field of Winkler and von Platten (1961). The observed compositional scatter of the silicic rocks is significantly larger than the variations in sequences of equivalent recent rocks, suggesting, together with the petrography, that some major elements, particularly K and Na, were affected by alteration processes. Likewise, several incompatible trace elements have a wide range of concentrations in the rhyolites. Elements such as Rb, U and Th show a positive correlation with K. The relatively immobile elements such as La, Ce, Zr and Nb are intercorrelated, but generally do not have a clear relationship with mobile trace elements. The REE patterns of the rhyolites from the Antigonish Highlands are highly fractionated, with strong LREE enrichments (Fig. 39b). The relatively flat normalized multielement pattern (Fig. 42) and high concentrations of high field strength elements (Nb to Y on Fig. 42) relative to ORG are consistent with a within-plate or attenuated crust (i.e. rift) tectonic environment (Pearce et al., 1984a).

The two most commonly invoked models for the origin of the felsic volcanic rocks are fractional crystallization of basaltic or andesitic magma, and partial melting of crustal rocks associated with the rising basaltic magma. Both of these processes can account for the genetic relationship suggested by the interbedding of mafic and felsic rocks. The available data cannot reliably discriminate between these two models. The lack of intermediate rocks, the high proportions of felsic to mafic volcanic rocks and the relatively low concentrations of incompatible trace elements in the rhyolites as compared to the basalts (Tables 4 and 5) argue against the derivation of the felsic rocks by fractional crystallization. The anatectic model appears to best account for the observed compositions of the Cambrian rhyolitic volcanic rocks.

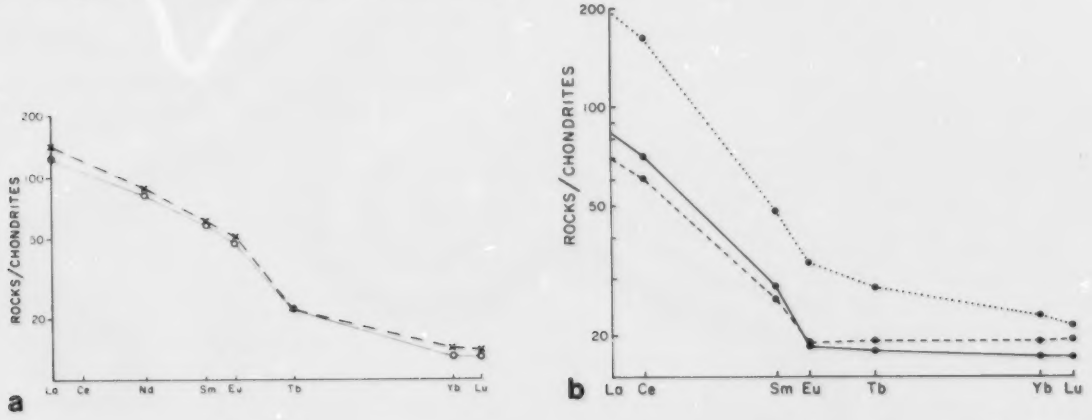


Figure 39. Chondrite-normalized REE abundances for the (**a**) mafic and (**b**) felsic volcanic rocks of the Arbuckle Brook Formation. Two mafic rocks analyzed are given different symbols for clarity.

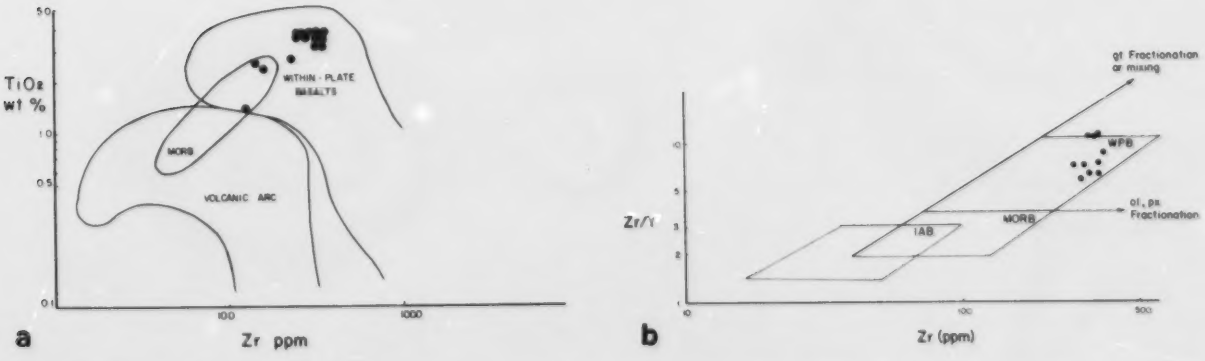


Figure 40. Tectonic setting of the mafic volcanic rocks of the Arbuckle Brook Formation. (**a**) after Pearce et al. (1981), (**b**) after Pearce and Norry (1979). Abbreviations as in Figure 20c.

Discussion

There are two general tectonic models for the Cambrian rocks of the Avalon Zone: a volcanic arc setting (Helmstaedt and Tella, 1973) and a rift setting (King, 1980; Keppie, 1982a,b, 1985b; Greenhough et al., 1985). Helmstaedt and Tella (1973) postulated that the volcanic rocks of the Middle Cambrian Bourinot Group in Cape Breton Island (Fig. 37) were calc-alkalic and formed part of a volcanic arc. They based this hypothesis upon the major element geochemistry of the rocks. However, it has been subsequently shown (Cameron, 1980 and herein) that major elements have experienced considerable mobility during alteration. The constraints upon tectonic modelling provided by the geochemistry of the Cambrian volcanic rocks therefore suggest a within-plate rift setting. The results presented here are consistent with those reported by Greenhough et al. (1985) for similar Cambrian volcanic rocks in the Avalon Zone of Newfoundland and southern New Brunswick. The relatively small volume of volcanic rocks suggests that this rifting was minor and therefore was probably not related to the opening of a major ocean (e.g. Iapetus). This is also borne out by the fact that the dyke swarms in the ancient North America craton related to the birth of Iapetus are older (i.e. 615 ± 10 Ma, Stukas and Reynolds, 1974).

Minor rifting in a within-plate tectonic setting is consistent with the tectonic model proposed by Keppie (1982a, b, 1985b; Murphy and Nance, 1989), in which the Cambrian basins in the Avalon Zone in the Maritimes represent pull-apart basins associated with dextral strike-slip faulting. Active faulting persisted through Middle Cambrian times, as indicated by the contemporaneous volcanism, generally waning in Late Cambrian to earliest Ordovician time with the subsequent cooling and subsidence of the basins. This faulting was inferred to have taken place during a period of transpression during the latest stages of the late Precambrian Avalonian-Cadomian Orogeny (Keppie, 1982a, 1985b; Keppie and Dostal, in press) when Precambrian rocks, interpreted as part of a volcanic arc-interarc basin complex, were obliquely telescoped. The Cambrian pull-apart basins appear to be generally coincident in space with Precambrian basins, suggesting some degree of inheritance (Keppie, 1982b). In the Avalon Zone in Newfoundland, the Cambrian rift volcanic rocks rest upon Proterozoic rocks containing mainly rift-related volcanic rocks, although limited volcanic-arc rocks are also present (Strong et al., 1978).

Structure and metamorphism

Minor structures

D_1 structures

The earliest structures observed in the Cambro-Ordovician rocks include isoclinal folds, associated thrusts and an axial planar cleavage which varies from a penetrative slaty to fracture cleavage and is generally parallel to the bedding in the fold limbs. These structures are best developed and are most complex in the upper parts of the Cambro-Ordovician sequence and in the basin centre, and die out stratigraphically downward and toward the ends of the basin (Fig. 31). Thus, at the southwestern end of the basin, a transition may be observed in west Iron Brook from the undeformed Black John Formation to minor thrusts and local reversals of facing directions in the Little Hollow Formation. Minor F_1 folds were recorded at only two places: in MacKinnons Brook and in East Doctors Brook. Here, the folds are isoclinal and have an S_1 axial planar cleavage defined by chlorite, muscovite and, locally, minor biotite. The presence of these minerals indicates an associated greenschist facies metamorphism. The mafic rocks of the McDonalds Brook Group contain the assemblage albite-quartz-epidote-chlorite-calcite-sphene ± magnetite ± rutile. This assemblage is consistent with greenschist facies metamorphism.

D_2 structures

Structures attributed to this set are best developed around the central part of the Cambro-Ordovician basin. They include folds and an axial planar fracture cleavage, which, although variable in orientation due to later folding, generally trends north-northwest-south-southeast (Fig. 31). The minor F_2 folds occur only in the pelitic rocks and are generally chevron-kink style folds which deform the S_1 cleavage. The lack of metamorphic minerals associated with these structures indicates a low, subgreenschist facies of metamorphism.

D_3 structures

These structures are also most common in the centre of the basin and are consistently oriented in a northeast-southwest plane across the area. They include ubiquitous minor folds and northwest-southeast steeply dipping axial planar fracture cleavage. The minor folds are generally upright-asymmetrical, open-close and chevron to concentric in style. They clearly deform both bedding and the S_1 cleavage in the Cambro-Ordovician rocks (Fig. 43). Their relationship with

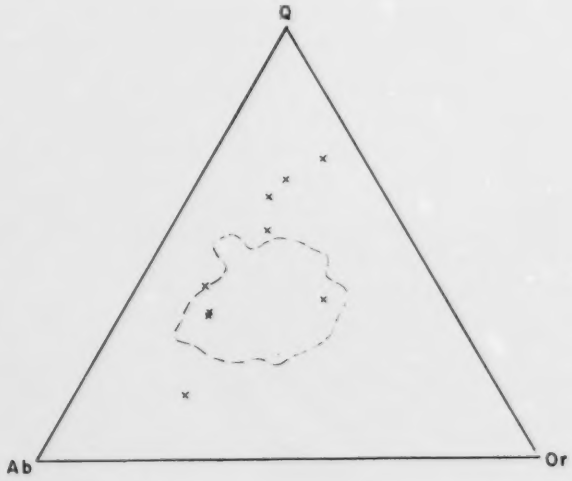


Figure 41. Ab-Or-Q normative diagram for the felsic volcanic rocks of the Arbuckle Brook Formation. Field of granitic rocks (after Winkler and von Platten, 1961) is also shown (dashed line).

the F_2 folds may be deduced from the major interference patterns discussed below. The lack of any significant recrystallization in the S_3 cleavage indicates high level subgreenschist facies metamorphic conditions during this folding event.

It is important to note that a steeply dipping northeast-southwest slaty fracture cleavage and associated minor folds represent the only set of structures recorded in the Silurian rocks in Iron Brook. These structures are correlated with the D_3 structures in the Cambro-Ordovician rocks. They are parallel to the major northeast-southwest syncline recorded north of the Hollow Fault (Benson, 1974; Boucot et al., 1974).

Major structures

Detailed mapping of two areas critical to the determination of the major structure was undertaken (Fig. 44a,b). The MacKinnons Brook area (Fig. 44a) straddles the hinge of the major F_1 fold in the eastern half of the area (Fig. 31). The outcrop pattern and structural data indicate an interference pattern in which a planar northeast-southwest S_3 cleavage parallel to a major tight F_3 fold (and minor F_3 fold axes) deforms a pre-existing tight-isoclinal fold. Structural readings around the axial trace reveal that the early fold closes downward while way-up structures indicate that the rocks become younger away from the axial trace. This is explicable in terms of an overturned anticline (i.e. a synformal anticline). Bedding-S_1 relationships in the hinge zone of the major early

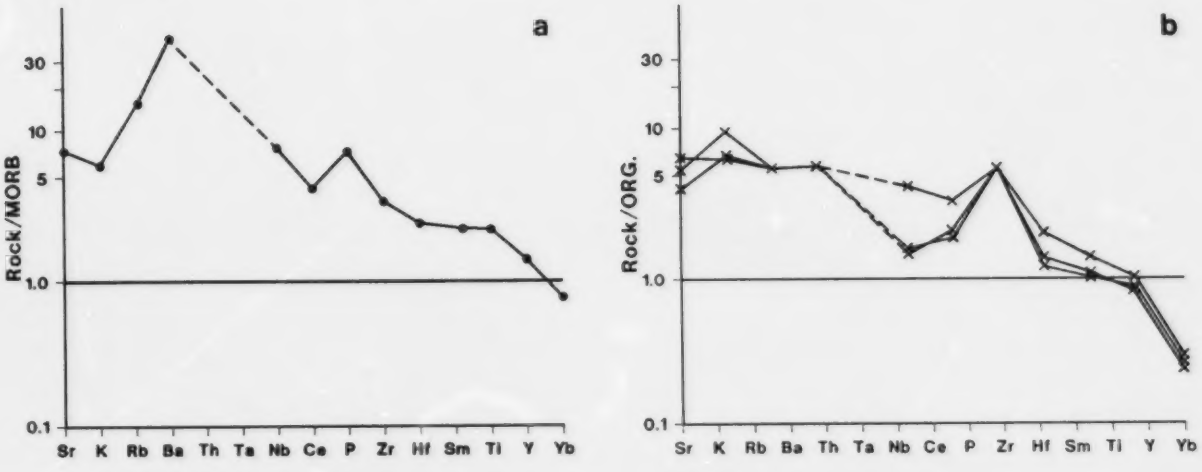


Figure 42. Multielement diagrams showing (**a**) the average MORB-normalized abundances of the mafic volcanic rocks and (**b**) ORG-normalized abundances for the felsic volcanic rocks of the Arbuckle Brook Formation (after Sun et al., 1979; Pearce et al., 1984a,b).

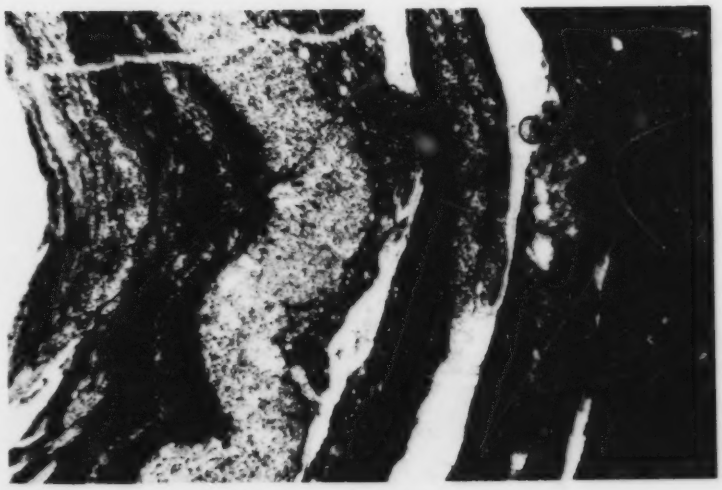


Figure 43. Photomicrograph (4 mm wide) showing structural relationships in the Black John Formation, MacKinnons Brook area. S_0 (bedding) and S_1 (cleavage) are parallel and are folded about an F_3 hinge (see text). A moderately developed fracture cleavage is axial planar to the F_3 fold.

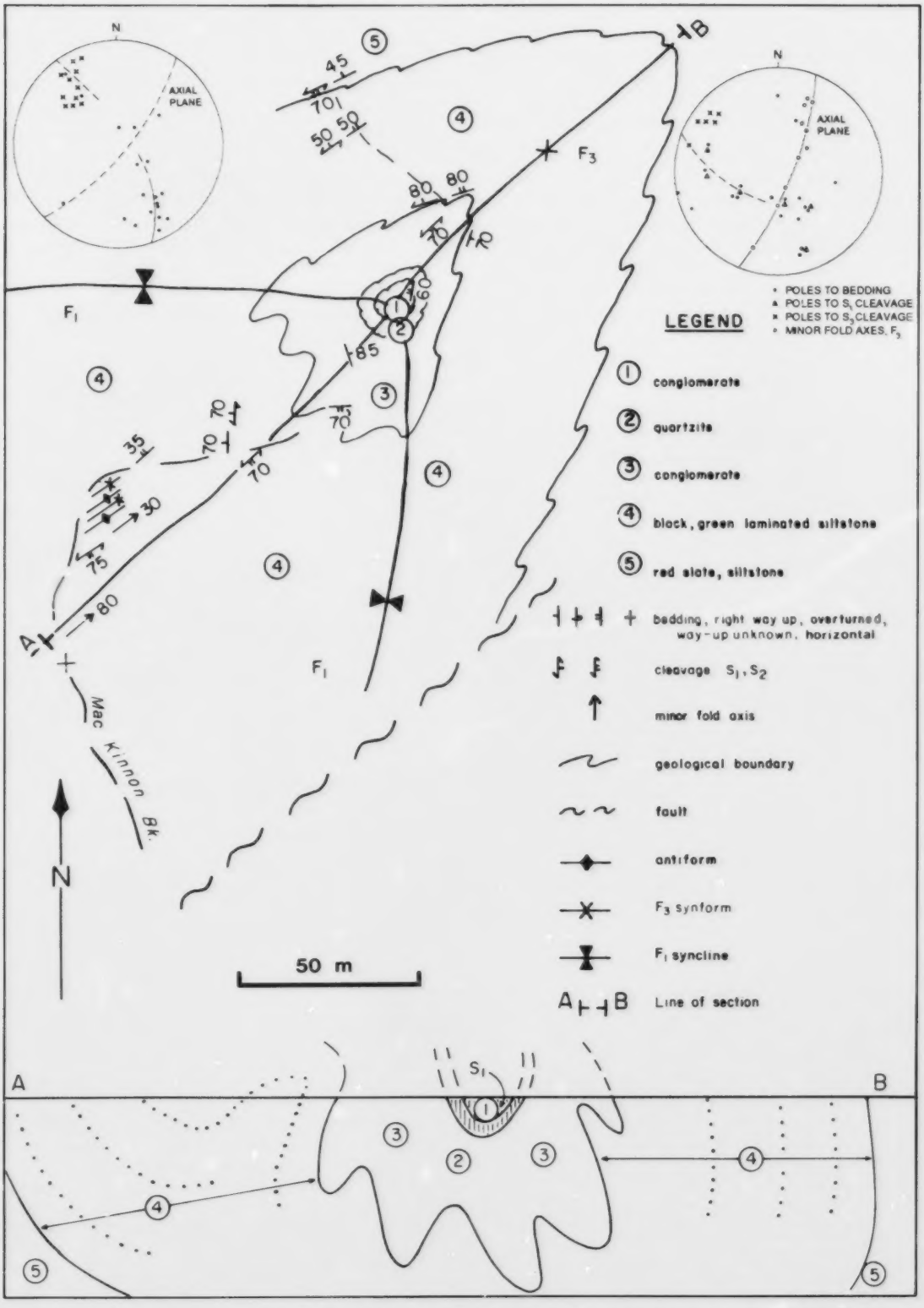


Figure 44a. Geological map of the MacKinnons Brook area. The map shows the distribution of the lithologies in northern MacKinnons Brook, structural data and the cross section A-B.

fold indicate that it is an F_1 fold. These relationships are revealed on the A-B cross section (Fig. 44a). The planar distribution of F_3 minor fold axes and S_0/S_3 intersection lineations is explicable in terms of superposition upon previously folded S_0 surfaces.

In the East Doctors Brook area (Fig. 44b), detailed mapping reveals that the Arbuckle Brook Formation has been thrust over the Iron Brook Group and folded by a northwest-southeast-trending anticline. All of these structures were subsequently modified by northeast-southwest, steeply inclined F_3 folds, separated by a thrust in the Iron Brook Group. These relationships are revealed in the two cross sections (Fig. 44b).

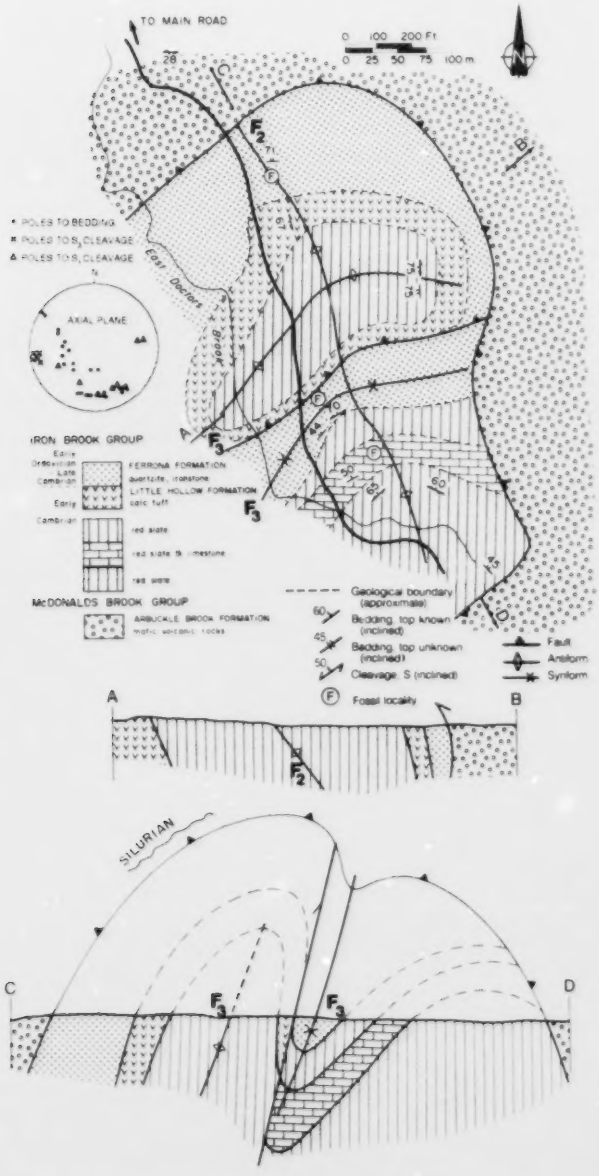


Figure 44b. Geological map of the East Branch Doctors Brook area with structural data on a stereoplot and cross sections A-B and C-D.

The minor and mesoscopic structures described above are consistent with the major structure revealed by the outcrop pattern on Figure 31. Thus, two northeast-southwest F_3 fold traces traverse the area (Fig. 31, cross section C-D). These deform earlier approximately orthogonal fold traces and thrusts. These earlier folds and thrusts are best observed on the cross section A-B parallel to an F_3 axial trace because the effect of F_3 folding is essentially absent. This cross section assumes that there is only one facies change between the McDonalds Brook and Iron Brook groups (Fig. 31). In cross section A-B, the structure is revealed as a major northeast-vergent recumbent fold of the Iron Brook Group thrust over the McDonalds Brook Group, modified by southeast-vergent back-thrusting, and refolded by northwest-southeast folds verging away from a central suture zone, as the basin closed. The structure is relatively simple at the northeast and southwest ends of the basin. The inferred sequence of events is depicted in Figure 31b. The sub-Silurian unconformity cuts across the F_1 and F_2 structures, thereby providing an upper limit on their age.

Cross sections C-D and E-F show the final stage of the deformation in which northeast-southwest folds are superimposed on the pre-existing structures.

On a regional scale, the major F_3 folds appear to be localized adjacent to the Hollow Fault. They also consistently trend anticlockwise of the Hollow Fault.

Age of structures

The lack of D_1 and D_2 structures in the Silurian rocks brackets their time of deformation between the Early Ordovician age of the Ferrona Formation and the Early Silurian age of the Beechill Cove Formation. The presence of D_3 structures in the Silurian-Early Devonian rocks north of the Hollow Fault indicates that these structures are considerably younger. North of the Hollow Fault, deformed Early Devonian rocks of the Knoydart Formation are unconformably overlain by tilted rocks of the Middle Devonian McAras Brook Formation (Keppie et al., 1978). This tightly confines the time of D_3 deformation to the Early-early Middle Devonian.

Origin of structures

The northwest-southeast orientations of both the F_1 recumbent folds and the F_2 upright folds, the decrease in structural complexity away from the centre of the Cambro-Ordovician basin and the opposite vergence of F_1 folds about the central axis, all point toward a central northwest-southeast-trending root zone acting as a locus for the deformation. It is postulated that the D_1 and D_2 structures were produced by one period of progressive deformation during closure of the basin. An early stage of northeast overthrusting was immediately followed by a backthrusting refolding stage (Fig. 31b). Such movements were localized in and telescoped the

Cambro- Ordovican pull-apart basin about its northwest-southeast axis. These structures may have been produced by sinistral movements on the northeast-southwest faults bounding the pull-apart basin. Such localized movements would account for the limited extent of the D_1 and D_2 deformation.

The major F_3 Acadian fold traces trend obliquely north-northeast-south-southwest across the Cambro-Ordovician basin and the adjacent Silurian rocks north of the Hollow Fault (Fig. 31). It is postulated that these folds formed during dextral transcurrent movements on the northeast-southwest faults. This explains their apparent localization adjacent to these faults.

Conclusions

Cambrian-Lower Ordovician rocks occur only in the northernmost Antigonish Highlands where they unconformably overlie the Precambrian Georgeville Group. They are divided into two groups that are lateral facies equivalents. The Iron Brook Group is predominantly sedimentary consisting of red clastic sediments, pink limestones containing late Early Cambrian fossils, quartzites, calcareous tuffs and ironstones containing Late Cambrian-Ordovician inarticulate brachiopods. The McDonalds Brook Group is mainly volcanic and consists of red clastic rocks at the base overlain by bimodal alkalic continental rift volcanic rocks with minor interlayered red clastic rocks and late Early Cambrian fossiliferous pink limestones.

Considered as a whole, the Cambro-Ordovician stratigraphy illustrates the progressive development of a small rift basin. Initially, terrestrial sediments deposited on a proximal braidplain (Malignant Cove Formation) eroded from mountainous areas possibly bounded by active faults (Fig. 45). Continued rifting led to the eruption of bimodal volcanic rocks initially into a braidplain-braided river environment (the Black John Formation and the lower member of the Arbuckle Brook Formation), and subsequently into a shallow marine environment (the Little Hollow Formation and the upper member of the Arbuckle Brook Formation). The volcanism is attributed to the formation of a pull-apart basin with dextral movement on the northeast-southwest bounding faults inducing northwest-southeast rifting which formed the locus of igneous activity (Fig. 45). The absence of tholeiitic plateau basalts and the restricted occurrence of Cambrian volcanic rocks in Nova Scotia suggests that this rift basin did not develop into a true ocean basin. However, the presence of cosmopolitan trilobites indicates that this rift basin was connected to the large oceans. The absence of Middle to Upper Cambrian rocks may be the result of either nondeposition, or deposition and subsequent erosion. A further shallow marine incursion is indicated in the Early Ordovician and was possibly due to renewed rifting leading to further volcanism (Ferrona Formation).

The limited extent of these Cambro-Ordovician rocks is believed to be controlled by the fault-bounded nature of the basin (Fig. 31). It is postulated that this basin formed as a pull-apart basin produced by dextral movements on the bounding northeast-southwest faults. The locus of rifting occurred where the Greendale Fault curves into the Hollow Fault. The geochemistry of the volcanic rocks is consistent with a continental transtensional tectonic setting. The northwest-southeast trend of the associated Cambrian gabbro bodies may be parallel to the rift axis.

During the Ordovician, there was a reversal of the movement on the bounding faults which became sinistral. The pull-apart basin became the locus of local compression inducing basin inversion as the central basin rocks were thrust out onto the margins (Fig. 45). Cambro-Ordovician rocks in the central part of the basin were thrust northeastward followed by back thrusting associated with northwest-southeast folds verging outward away from a central suture zone. This complex structure dies out toward the northwestern and southwestern ends of the basin, where the oldest Cambrian units and the underlying Precambrian rocks are relatively mildly deformed and the basal unconformity between Cambro-Ordovician and Precambrian rocks is preserved. These structures are inferred to have been produced by a reversal of motion on the northeast-southwest bounding faults to a

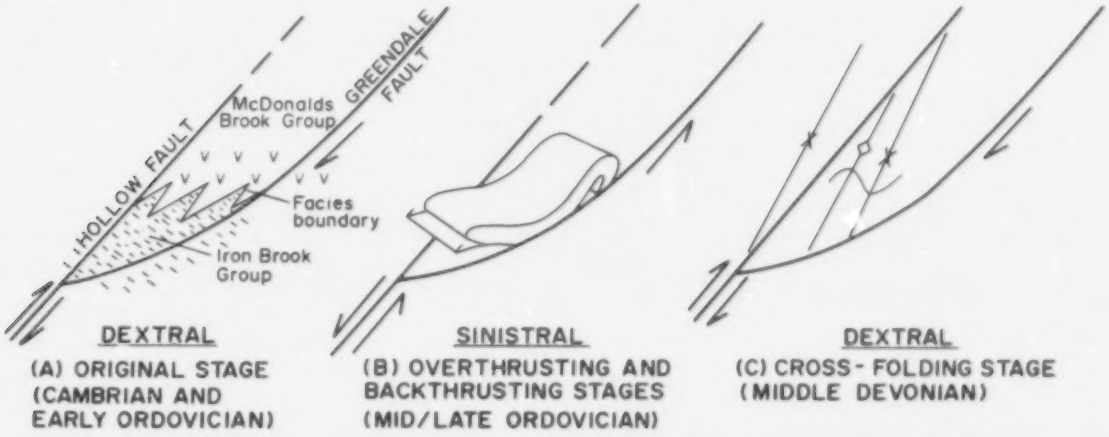


Figure 45. Schematic structural diagram showing the development (A) and telescoping (B) of the pull-apart basin. The sequence was subsequently deformed by a cross-folding stage (C).

sinistral sense. These movements expelled the Cambro-Ordovician rocks lying on attenuated lithosphere from the basin centre onto the basin margins. This accounts for the limited extent of the deformation. These structures are truncated by the sub-Silurian unconformity, thereby constraining the time of deformation to the Ordovician. Both the Cambro-Ordovician and Siluro-Devonian rocks were deformed during the Devonian by upright northeast-southwest folds that were preferentially developed adjacent to the major northeast-southwest faults. The folds are consistently oriented slightly anticlockwise of the faults thereby indicating renewed dextral movements on the northeast-southwest faults (Fig. 45).

LATE ORDOVICIAN-EARLY DEVONIAN ROCKS

Stratigraphy

Definitions of previous workers

Upper Ordovician-Lower Devonian rocks in the Antigonish Highlands are assigned to the Arisaig Group and have been studied in detail by Maehl (1961) and Benson (1974) and to the north of the Hollow Fault by Boucot et al. (1974). They are widely distributed throughout the Antigonish Highlands and unconformably overlie the late Precambrian Georgeville Group and the Cambrian-Lower Ordovician Iron Brook and McDonalds Brook groups.

A summary of the stratigraphy proposed by previous workers is given in Table 6. Maehl (1961) divided the Arisaig Group into two distinct sequences (the northern and southern highlands) which he interpreted to be lateral facies equivalents. In the northern highlands, he interpreted the Bears Brook Formation to be disconformably overlain by the Arisaig Group and to conformably overlie the Cambro-Ordovician Browns Mountain Group (Table 6). In the southern highlands, he defined the Charcoal, Sunnybrae, Glencoe Brook and Kerrowgare formations. The Charcoal Formation was thought to compose the upper part of the Browns Mountain Group, the Sunnybrae Formation was interpreted to be a lateral equivalent to the Bears Brook Formation, the Glencoe Brook Formation was considered to be a lateral equivalent to the Beechill Cove Formation and the Kerrowgare Formation was interpreted to be a lateral equivalent to the Ross Brook, French River and McAdam formations (Table 6).

The Charcoal Formation was thought to consist of 600 m of felsic and andesitic volcanic rocks overlain by 350 m of dark grey-green banded shales and slates (Sunnybrae Formation), 300 m of quartzites, sandstones and mudstones (Glencoe Brook Formation) and by 200 m of blue-grey and green

Table 6. Comparative stratigraphic interpretations of the Arisaig Group

	Boucot et al., 1974	Maehl, 1961		Benson, 1974	This study
Devono-Carboniferous	Arisaig Group	Northern Antigonish Highlands	Southern Antigonish Highlands	Antigonish Highlands	Antigonish Highlands
	Undivided				
Lower Devonian	Arisaig Group: Knoydart Fm	Knoydart Fm	Knoydart Fm	Knoydart Fm	Knoydart Fm
	Stonehouse Fm	Stonehouse Fm	Stonehouse Fm	Stonehouse Fm	Stonehouse Fm
Upper Silurian	Moydart Fm	Moydart Fm		Moydart Fm	Moydart Fm
	McAdam Brook Fm	McAdam Fm	Kerrowgare Fm	McAdam Fm	McAdam Fm
	Doctors Brook Fm				
	French River Fm	French River Fm		French River Fm	French River Fm
Lower Silurian	Ross Brook Fm	Ross Brook Fm		Ross Brook Fm	Ross Brook Fm
	Beechill Cove Fm	Beechill Cove Fm	Glencoe Brook Fm	Beechill Cove Fm	Beechill Cove Fm
Upper Ordovician	Bears Brook Volcanic Group	Bears Brook Fm	Sunnybrae Fm	Dunn Point Volcanics	Bears Brook Fm
Cambrian - Lower Ordovician	Browns Mountain Group	Browns Mountain Group	Charcoal Fm / Browns Mountain Group	Browns Mountain Group	Iron Brook Group; McDonalds Brook Group
Precambrian					Georgeville Group

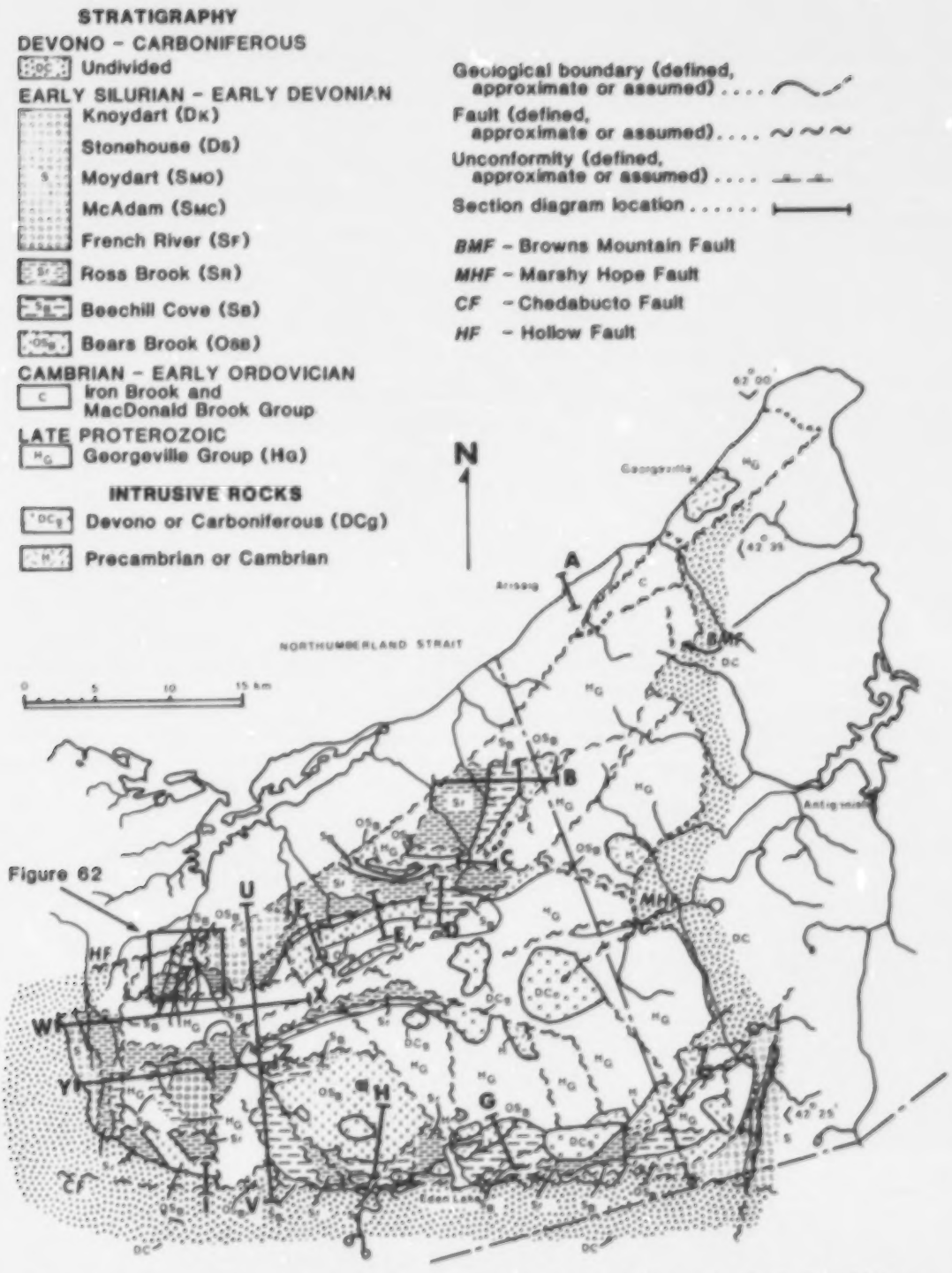


Figure 46. Generalized geological map of the Antigonish Highlands showing the distribution of Upper Ordovician-Silurian and plutonic rocks (see Map 1749A for details). For details of Devonian-Carboniferous rocks see Boehner and Giles (1982) and Yeo and Gao (1986). Section diagrams of locations A to I are shown in Figure 48. Inset shows location of the detailed geology of the McLellans Brook area (Fig. 62).

siltstones, shales and slates (Kerrowgare Formation). These formations were interpreted by Maehl (1961) to be overlain by the Stonehouse and Knoydart formations.

Remapping of the western Antigonish Highlands (Murphy, 1986) indicates that the definitions of these formations are redundant because they are very similar to rocks of equivalent age in the northern Antigonish Highlands. Thus it is proposed here that the names of these formations should be abandoned and replaced by their better known northern equivalents. The rocks assigned by Maehl (1961) to the Charcoal Formation are similar in lithology and in stratigraphic position to the Bears Brook Formation. The dark grey-green shales of the Sunnybrae Formation are generally similar to those of the Ross Brook Formation. This interpretation is supported by Benson (1967) who reported fossils typical of the Ross Brook Formation within the Sunnybrae Formation as defined by Maehl (1961). Therefore, the contact between the Charcoal and Sunnybrae formations in the southern highlands is reinterpreted herein as the overstepping of the Ross Brook Formation onto the Bears Brook Formation. In general the rocks assigned by Maehl (1961) to the Glencoe Brook Formation strongly resemble Beechill Cove facies 4 and 5 of Pickerill and Hurst (1983) in the type section of the Beechill Cove Formation. Furthermore, *Leptostrophia beechillensis*, typical of the Beechill Cove Formation has been identified (R. Pickerill, pers. comm., 1985) from a sequence assigned by Maehl (1961) to the Glencoe Brook Formation. The Kerrowgare Formation lithologically resembles the French River and McAdam formations and similar fossils have been reported from there by Benson (1967). It is therefore recommended that the Charcoal, Sunnybrae, Glencoe Brook and Kerrowgare formations defined by Maehl (1961) in the southern Antigonish Highlands are abandoned because although there are some facies variations between the southern highlands and the type area, the definitions of the formations in the type area can be applied to these rocks. The main difficulty in applying the definitions of the type area to the southern highlands lies in the relatively poor quality of outcrop and in the intense localized deformation rather than in primary facies variations as envisaged by Maehl (1961).

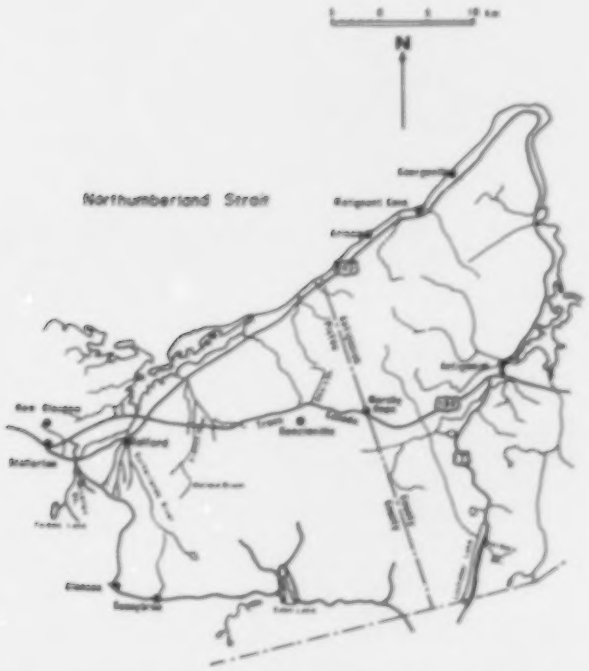


Figure 47. Place names and type localities mentioned in the text.

Benson (1967, 1974) proposed a similar stratigraphic sequence to that of Maehl (1961) for the Arisaig Group in the northern Antigonish Highlands (Table 6). The main distinction between these authors occurs in the proposed stratigraphy of the southern Antigonish Highlands. Benson (1974) did not distinguish between sequences of the northern and southern highlands (Table 6). For reasons outlined above, we agree with this interpretation.

Another important distinction between the sequences proposed by these authors concerns the stratigraphic position of the mafic to felsic volcanic rocks that underlie the Silurian fossiliferous siliciclastic rocks in the northern highlands. Benson (1974) assigned the volcanic rocks to the Cambro-Ordovician Brierly Brook Formation of the Browns Mountain Group and the Dunn Point Volcanics, whereas Maehl (1961) included the volcanic rocks in the Bears Brook Formation (Table 6). We agree with the interpretation of Maehl (1961). Mapping (Murphy et al., 1982; Murphy, 1986) shows that the Brierly Brook Formation as defined by Benson (1974) includes rocks that range from late Precambrian to earliest Silurian in age and therefore the definition of this formation is in error. To the north of the Hollow Fault, Benson (1974) assigns a sequence of felsic and mafic volcanic rocks that disconformably underlie the Silurian siliciclastic rocks (which he calls the Dunn Point Volcanics) to the uppermost Ordovician (Table 6). In this report, these rocks are interpreted as lateral equivalents to the Bears Brook Formation.

In general, the stratigraphy proposed by Maehl (1961) for the northern Antigonish Highlands closely corresponds with that of the type area of the Arisaig Group as defined by Boucot et al. (1974). However, the main difference between these authors lies in the interpretation of the sequence beneath the Beechill Cove Formation. In the Arisaig area, Boucot et al. (1974) redefined and upgraded the Bears Brook Formation of Maehl (1961) to group status (the Bears Brook Volcanic Group, Table 6) and subdivided this group into three formations: the Dunn Point Formation, which is overlain by the McGillivray Brook Formation, and the Malignant Cove Formation, which is laterally equivalent to both of these formations. Boucot et al. (1974) interpret the Bears Brook Volcanic Group to be disconformably overlain by the Beechill Cove Formation of the Arisaig Group and to unconformably overlie the Browns Mountain Group. However, on the basis of regional mapping in the northern Antigonish Highlands, Murphy et al. (1982) assigned the Malignant Cove Formation to the McDonalds Brook Group and interpreted it to be Early Cambrian in age. Therefore it cannot be a lateral equivalent to the Dunn Point and McGillivray Brook formations. In this study, the term Bears Brook Formation is retained in preference to the Bears Brook Volcanic Group because it is

argued (below) that the Bears Brook Formation forms part of a cogenetic sequence with the overlying formations of the Arisaig Group, and therefore should be considered part of the Arisaig Group (Table 6).

Another minor difference between these authors lies in the spelling of the Beechill Cove Formation (Table 6). Maehl's (1961) spelling is the same as the geographic place name and will be adopted in this paper.

Boucot et al. (1974) subdivided the McAdam Formation of Maehl (1961) into the Doctors Brook and McAdam Brook formations. The contact between these formations is placed at the base of the black shales of the McAdam Brook Formation, which overlie grey shales and thin (up to 5 cm thick) siltstone beds of the Doctors Brook Formation. However, this distinction could not be made in the Antigonish Highlands. It is therefore proposed that all three formation names should be retained, and that the McAdam Brook and Doctors Brook formations in the Arisaig area should be viewed as lateral facies variations of the McAdam Formation in the Antigonish Highlands.

Age

In general, the lithology, stratigraphy and fauna of the Arisaig Group in the Antigonish Highlands are very similar to the type area of the Arisaig Group to the north of Hollow Fault (Boucot et al., 1974), and correlations have been proposed by Maehl (1961) and Benson (1974). The extrapolation of these formations across Hollow Fault is based on lithological, stratigraphic, paleontological and geochemical similarities. The distribution of the formations of the Arisaig Group is shown in Figure 46 and the place names and type localities mentioned in the text are shown in Figure 47. With the exception of the Bears Brook Formation, the age of these rocks is based on paleontological evidence (Maehl, 1961; Benson, 1974; Boucot et al., 1974; Smith, 1979). Thus the Beechill Cove and Ross Brook formations are assigned to the Early Silurian, the French River, McAdam and Moydart formations to the Late Silurian; and the Stonehouse and Knoydart formations to the Early Devonian.

The age of the Bears Brook Formation is less precisely known. On the basis of the above stratigraphic relationships, it lies between strata of late Precambrian and Early Silurian age. To the north of the Antigonish Highlands, mafic and felsic volcanic rocks underlying the Beechill Cove Formation have been assigned to the Dunn Point and McGillivray Brook formations. Rb-Sr studies of the felsic volcanic rocks in these formations have yielded an age of 421 ± 11 Ma (Fullager and Bottino, 1968; recalculated in Keppie and Smith, 1978) which together with the stratigraphic context suggests that these rocks are uppermost Ordovician or lowermost Silurian in age.

Recent Rb-Sr analyses of the Bears Brook Formation show abundant scatter yielding an errorchron of 425 ± 37 Ma (R.F. Cormier, pers. comm., 1986). However, when taken in conjunction with the stratigraphic context and lithological association (i.e. these rocks underlie the Lower Silurian Beechill Cove Formation), it is probable that the Bears Brook Formation of Maehl (1961) is coeval with the Dunn Point and McGillivray Brook formations and is therefore uppermost Ordovician or lowermost Silurian in age.

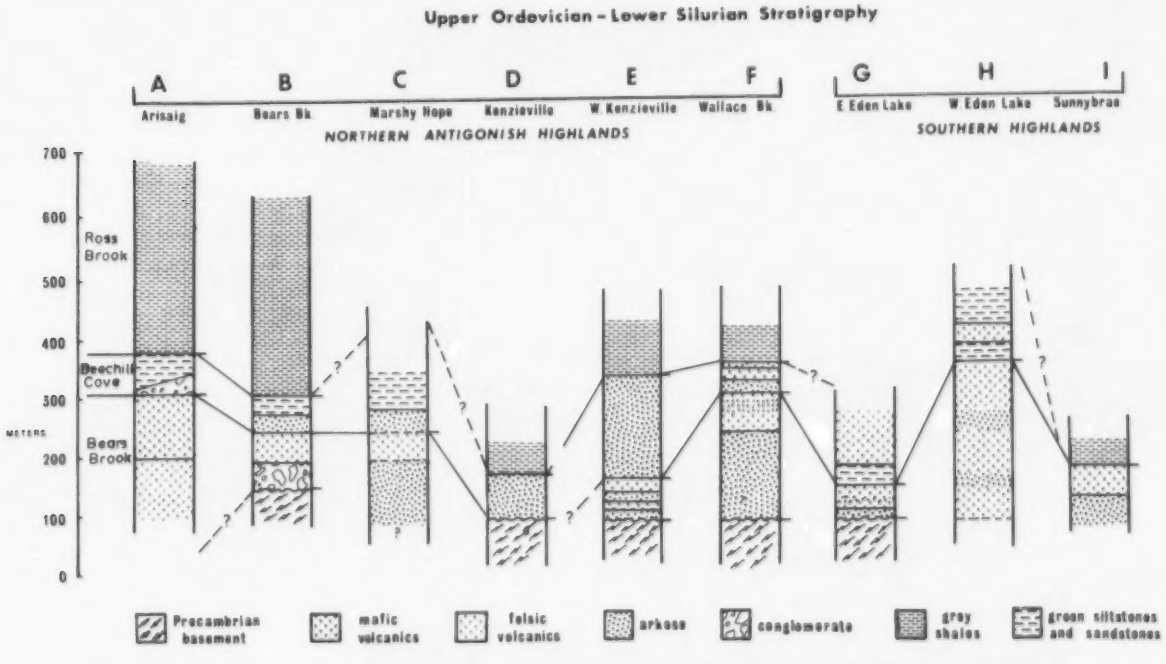


Figure 48. Stratigraphy of Upper Ordovician-Silurian rocks. Section A is generalized after Boucot et al. (1974), Keppie et al. (1978), Pickerill and Hurst (1983) and Hurst and Pickerill (1986). Locations of sections

Bears Brook Formation

Field relationships. The Bears Brook Formation of Maehl (1961) consists of red conglomerate, arkose, shale and interlayered mafic and felsic volcanic rocks. This formation unconformably overlies the late Precambrian Georgeville Group and is conformably or disconformably overlain by arkoses and fossiliferous blue-green siltstones of the Beechill Cove Formation. The top of the Bears Brook Formation is placed at the uppermost volcanic rock that is stratigraphically beneath the fossiliferous siltstones. This definition coincides with that of the base of the Beechill Cove Formation in its type area (Boucot et al., 1974). The type section of the Bears Brook Formation is on Wallace Brook (Fig. 47; see also Maehl, 1961). The formation is not exposed in Bears Brook; however, good sections are exposed on logging roads to the north and west of Bears Brook and west of Eden Lake. In the latter area, vertical gradient magnetic maps (Geological Survey of Canada, 1982; see also Figure 10) and radar imagery (see frontispiece) have a concentric pattern, consistent with the presence of a vent complex.

The Bears Brook Formation shows significant lithological variations across the fault blocks. For example, to the north of the Browns Mountain Fault (e.g. sections D and E, Fig. 48), arkoses and conglomerates predominate and volcanic rocks are relatively minor. To the south of this fault, red beds are interlayered with mafic and felsic volcanic rocks in the western highlands whereas in the southern highlands felsic volcanic rocks predominate over mafic flows and red beds (section H, Fig. 48). Correlations across fault blocks are based on the similarity of the depositional environment and the stratigraphic position of the Bears Brook Formation between the Georgeville Group and the Beechill Cove Formation. The best exposures of the red beds of the Bears Brook Formation occur on the Trans-Canada Highway in the Marshy Hope area. The mafic volcanic rocks and interlayered red beds are well exposed on Wallace Brook and in the Sunnybrae area. The felsic volcanic rocks are best exposed in the Eden Lake area.

The thickness of the Bears Brook Formation is highly variable, mainly due to the changing thickness of the volcanic rocks. In the northern highlands, the maximum thickness is about 200 m in the Wallace Brook area. In the southern highlands, the succession is at least 400 m thick.

Description. Conglomerates and arkoses are massive and typically red, but may contain large irregularly-shaped green-stained patches. The conglomerates are poorly sorted and contain red silty and sandy lenses. They are deposited in channels up to 3 m across. The arkoses are well sorted and are characterized by "rip-up" clasts of red shale up to 5 cm in length. They display channels up to 5 m across with many examples of normal and inverse grading. Shales are relatively subordinate. They consist of beds varying from 1 cm to 1 m in thickness. The thinner beds are commonly discontinous on an outcrop scale.

The conglomerate and the arkose are polymictic, containing subangular to subrounded clasts of red and green shale, mafic and felsic volcanic rocks, granite, diorite, quartz, albitized plagioclase and, more rarely, sericitized orthoclase and disseminated pyrite. The matrix consists of fine grained clastic material, hematite, sericite, chlorite and epidote. Carbonate is variable in abundance and is most common adjacent to faults. The shales contain subrounded to subangular quartz in a matrix of hematite, chlorite and sericite.

Mafic volcanic rocks are green, massive and fine grained. They commonly contain amygdules filled with quartz, calcite, zeolites and pumpellyite. They occur in flows up to 3 m thick. In thin section, mafic volcanic rocks consist of thin laths of saussuritized plagioclase (albite with epidote and calcite) displaying a flow texture. Some samples contain relict augite, partially altered to chlorite. Opaques (generally titaniferous magnetite) are highly variable in abundance and display both equant and interstitial textures. Apatite is spatially associated with magnetite. Alteration of these rocks has produced chlorite, epidote and sphene ± actinolite ± quartz ± calcite ± rutile.

Felsic volcanic rocks consist of ignimbrites, tuffs and flows. Lahars are spatially associated with the felsic volcanic rocks. These lithologies closely resemble those of the Precambrian felsic volcanic rocks of the Keppoch Formation, from which they can only be reliably distinguished by their stratigraphic position. Ignimbrites display pink to red banding that is approximately parallel to bedding. Good examples of banded ignimbrites occur in the Eden Lake area. Banding in the ignimbrites is defined in thin section by flattened pumice. They contain lithic and crystal fragments up to 5 cm in length. The matrix is variably welded and consists of flattened devitrified pumice and shards in a fine grained cryptocrystalline quartzofeldspathic matrix (Fig. 49).

Nonwelded ignimbrites contain well preserved tricuspate shards (Fig. 49). The lithic fragments are almost exclusively of felsic composition and are generally less than 4 mm in size. Quartz phenoclasts are commonly cracked and embayed. Plagioclase is commonly saussuritized and altered to albite. Orthoclase phenoclasts are partly sericitized.

Tuffs contain lithic and crystal fragments up to 5 cm in length supported in a fine grained ash matrix. They are massive, pale grey to red and fine grained, containing angular fragments up to 1 cm in length. Flows are massive, pale grey to pink and are generally aphyric, though they may contain up to 10 percent saussuritized plagioclase phenocrysts up to 8 mm in length set in a cryptocrystalline quartzofeldspathic matrix. Quartz phenocrysts, where present, are in part euhedral and in part embayed. The total amount of phenocrysts rarely exceeds 5 modal percent. Lahars consist of poorly sorted angular fragments of felsic and mafic volcanic rocks in a matrix dominated by chlorite and sericite.

Beechill Cove Formation

Field relationships. The Beechill Cove Formation can be subdivided into two members. The lower member consists of up to 80 m of fluviatile arkoses and conglomerates. The

upper member contains shallow-marine fossiliferous massive blue-green siltstones, minor rhyolites and grey laminated siltstones. In most sections, the contact between the two members is sharp; however, minor interlayering occurs in some sections. The type section of the Beechill Cove Formation is north of the Hollow Fault near Arisaig, but does not contain the lowermost member. Fossil assemblages establish an earliest Silurian age in the type area (Pickerill and Hurst, 1983) and similar assemblages occur in the Beechill Cove Formation in the Antigonish Highlands (Smith, 1979; R. Pickerill and J. Hurst, pers. comm., 1984), also indicating an Early Silurian age.

The Beechill Cove Formation conformably to disconformably overlies the Bears Brook Formation and conformably underlies the Ross Brook Formation. Conformable relationships occur between the lower member of the Beechill Cove Formation and the Bears Brook Formation. Disconformable relationships occur where the lower member is missing. The top of the Beechill Cove Formation is defined by the first appearance of black micaceous mudstones (see Boucot et al., 1974) of the lower member of the Ross Brook Formation.

The lower member red beds show considerable lateral variation in thickness. In some sections (e.g. the Trans-Canada Highway near Marshy Hope), they attain a thickness of about 80 m whereas in others the red beds are absent (e.g. section D, Fig. 48). The upper siltstone member commonly oversteps the Bears Brook Formation and rests unconformably on the Precambrian basement (Fig. 46, Map 1749A). Laminated siltstones predominantly occur in the southern highlands (e.g. in the Eden Lake area) where they disconformably overlie the Bears Brook Formation. This unit varies considerably in thickness. It is up to 20 m thick in the Eden Lake area; however, these rocks are commonly interlayered with a regular layered siltstone unit. The rhyolites are thickest in sections that also contain thick rhyodacites in the underlying Bears Brook Formation, and are best exposed in the Eden Lake area. These rhyolites are not placed in the Bears Brook Formation because they are interlayered with or stratigraphically overlie the fossiliferous siltstones that occur at the base of the Beechill Cove Formation.

Description. The conglomerates, arkoses and shales are virtually identical to those of the Bears Brook Formation. The distinction between the two formations in the field is based on their stratigraphic position with respect to the underlying felsic volcanic rocks of the Bears Brook Formation. The conglomerates and arkoses are red, display normal and inverse grading and are deposited in channels up to 5 m across. The siltstones are massive, blue-green in colour and contain abundant muscovite. In most localities, the blue-green siltstone contains abundant fossils in beds up to 30 cm in thickness. Weathered surfaces display characteristic limonitic staining. Fossils consist of brachiopods, trilobites, crinoids and corals (Smith, 1979; Pickerill and Hurst, 1983). The massive siltstones strongly resemble the "regular layered facies" (facies 4) of Pickerill and Hurst (1983) from the Beechill Cove Formation of the type area. The siltstones are laterally continuous and display sharp basal contacts. Fossils, where they occur, are found at the base of individual beds. The laminated siltstones resemble the "lenticular siltstone facies" (facies 5) of Pickerill and Hurst (1983) in that they are plane laminated with individual laminae (ranging up to 3 cm in thickness), they have sharp planar basal contacts and are laterally discontinuous. They occur only in the southern highlands.

Ross Brook Formation

Field relationships. The Ross Brook Formation conformably overlies the Beechill Cove Formation. However, in certain localities, the Ross Brook Formation oversteps the Beechill Cove Formation to rest directly on the Bears Brook Formation or the Precambrian basement. An excellent example of this occurs at Sunnybrae on Glencoe Brook where the Ross Brook Formation rests on the Bears Brook Formation (section I, Fig. 48). There is evidence of reworking of the Bears Brook Formation at this contact.

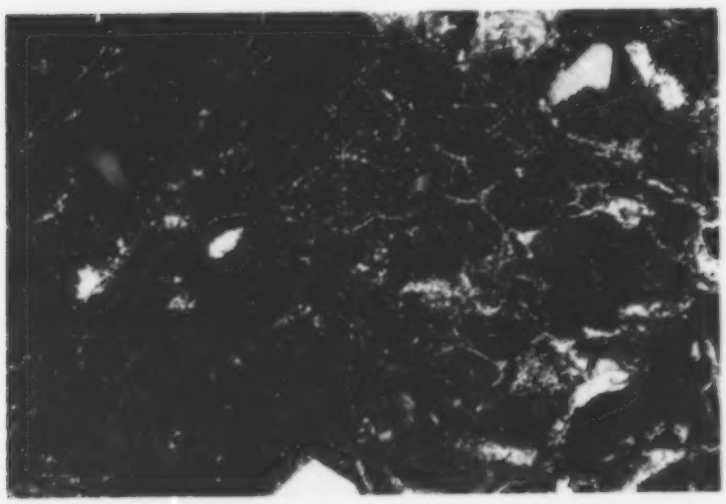


Figure 49. Photomicrograph (4 mm wide) of an ignimbrite, Bears Brook Formation. The ignimbrite contains phenoclasts of quartz (e.g. the lower part of the photograph) and plagioclase (now albite, upper right) in a cryptocrystalline quartzofeldspathic matrix. Note the well-preserved shards in the central-right part of the photograph.

The lower contact with the Beechill Cove Formation is placed at the first appearance of black or dark grey shales. The upper contact with the French River Formation is defined by a sharp colour and a lithological change from dark shales to pale green siltstones (Maehl, 1961; Boucot et al., 1974). The Ross Brook Formation is best exposed in the Kenzieville area, particularly on French River where it may be subdivided into lower and upper members based on the absence or presence (respectively) of massive sandstones or siltstones interbedded with the shales. The lower member varies from 30 to 100 m in thickness. The upper member is only exposed on French River, where it is about 100 m thick.

Description. The Ross Brook Formation consists of black to grey fossiliferous bioturbated micaceous shales (lower member) overlain by an interlayered sequence of grey-black to blue-grey shales and massive fossiliferous cross-bedded siltstones and sandstones (upper member; cf. Hurst and Pickerill, 1986).

The shales are poorly laminated and fissile. Weathered surfaces display limonitic staining. They are abundantly fossiliferous containing brachiopods, trilobites, crinoids, gastropods and, more rarely, graptolites (see Maehl, 1961; Benson, 1974). The sandstones and siltstones are grey, massive and resistant to erosion and display good cross-bedding. They generally weather to a pale brown or orange colour.

French River Formation

Field relationships. The French River Formation conformably overlies the Ross Brook Formation. The lower contact is defined by a sharp change in colour from the grey shales of the Ross Brook Formation to the paler sandstones and siltstones of the French River Formation. In the study area, the French River Formation is only identified on French River (the type section; see Maehl, 1961) and in limited exposures on Sutherlands River probably because it is difficult to distinguish from the overlying McAdam Formation in areas of poor exposure. The upper contact with the overlying McAdam Formation is gradational and is defined by the relative dominance of siltstones and shales over sandstones in the McAdam Formation. In the type section, the French River Formation is about 60 m thick.

Description. The French River Formation is dominated by blue-grey micaceous sandstones and siltstones with relatively subordinate shales. A set of fractures that are perpendicular to bedding give these rocks a blocky appearance. They contain abundant fossils, especially brachiopods and crinoids. Individual units are up to 30 cm in thickness and commonly display grading from a siltstone or sandstone base to a shaly top. Black concretions of mudstone occur near the base of the formation. The shales are commonly laminated, though the laminae are generally disturbed probably as a result of bioturbation. The rock is made up of angular to subangular quartz, albite and muscovite with relatively minor pyrite in a matrix consisting predominantly of sericite, chlorite and minor calcite.

McAdam Formation

Field relationships. The McAdam Formation conformably overlies the French River Formation. This contact is gradational and is placed where grey shales of the McAdam Formation predominate over the blue-grey sandstones and siltstones of the French River Formation. The upper contact with the overlying Moydart Formation is also gradational and is defined where green shales of the Moydart Formation predominate over the darker shales of the McAdam Formation. The thickness of the McAdam Formation in the study area is difficult to determine due to poor exposure. However, on Sutherlands Brook in the Telford area, the formation is estimated to be about 200 m thick. The McAdam Formation is also exposed on logging roads in the Forbes Lake area in the western Antigonish Highlands.

Description. The McAdam Formation is dominated by grey massive cross-bedded siltstone at the base with minor interbedded grey shale. The grey mudstone becomes dominant toward the middle of the formation. Individual siltstone units range up to 25 cm in thickness and resemble those of the Ross Brook Formation. However, as noted by Maehl (1961), the siltstones of the McAdam Formation may be distinguished from the Ross Brook Formation in that their abundance decreases rapidly up section. Furthermore, they are generally calcareous. Brachiopods and crinoids are abundant throughout this formation. The siltstones consist of well-sorted subangular to subrounded quartz, with relatively minor plagioclase and muscovite. The matrix consists of hematite, sericite, chlorite, calcite and clay minerals. The shales are similar in composition to the matrix of the siltstones.

Near the middle of the formation the grey shales are overlain by dark grey to black fissile shales and in turn by grey shales. The dark shales are characterized by the presence of calcareous nodules up to 5 cm in diameter.

Moydart Formation

Field relationships. The Moydart Formation may be divided into a lower green and an upper red member (Maehl, 1961). The lower contact with the underlying McAdam Formation is gradational and is placed where green-grey shales predominate over the darker shales of the McAdam Formation. The upper contact with the overlying Stonehouse Formation is conformable and is defined at the uppermost appearance of red shale. The Moydart Formation is best exposed in the Telford area, in fault slices on McLellans Brook and on the west side of Lochaber Lake. The thickness of the lower member is about 90 m, and the upper member is about 10 m thick.

Description. The lower member consists of green sandstones, siltstones and shales. The sandstones are massive, hard and micaceous. The siltstones are laminated, may display cross-bedding and commonly display graded bedding with shaly tops. Shales are the dominant lithology especially toward the top of the member. Ripple marks occur at the top of some of the siltstone units. The green member of this

formation contains abundant brachiopods and crinoids. The upper member consists of interbedded red, purple and green shales. They have a mottled appearance, are moderately calcareous and have calcareous nodules.

The siltstones and sandstones consist of between 35 and 60 percent subrounded to angular quartz with lesser amounts of plagioclase and muscovite in a matrix of sericite, chlorite, calcite and clay minerals. The shales are similar in composition to the matrix of the siltstones. The red shales also contain abundant hematite.

Stonehouse Formation

Field relationships. The Stonehouse Formation conformably overlies the Moydart Formation. The contact is defined at the uppermost red shale unit of the Moydart Formation. The upper contact is placed at the uppermost green unit that is conformably overlain by the red beds of the Knoydart Formation. The Stonehouse Formation is best exposed in the Forbes Lake area in the western Antigonish Highlands and on the west side of Lochaber Lake in the southeastern highlands. Thickness estimates are difficult due to incomplete exposure but in the Forbes Lake area, the Stonehouse Formation is probably at least 600 m thick.

Description. As noted by Maehl (1961), the lower part of the Stonehouse Formation resembles the green member of the Moydart Formation. These rocks are green-grey to blue-grey micaceous siltstones and commonly have well-developed laminated tops that commonly display cross-bedding. Blue-grey siltstones predominate in the middle of the formation, and toward the top of the sequence minor red shales occur. The Stonehouse Formation is very fossiliferous containing brachiopods and crinoids. Although Copeland (1964) reports ostracodes from this formation in the Arisaig area, none were positively identified in the Antigonish Highlands.

Knoydart Formation

Field relationships. The contacts between the Knoydart Formation and the underlying Stonehouse Formation and between the Knoydart Formation and the overlying McAras Brook Formation is not exposed in the area. However, Boucot et al. (1974) document a gradational contact between the Stonehouse and Knoydart formations in the Arisaig area. On McAras Brook, the Knoydart Formation is unconformably overlain by the McAras Brook Formation. The Knoydart Formation is best exposed near Lochaber Lake. The thickness of the formation is difficult to estimate due to the lack of exposure of the contacts but on Hurlbert Brook to the east of Lochaber Lake the section is at least 500 m thick. No fossils were found in the Knoydart Formation in the Antigonish Highlands.

Field description. The Knoydart Formation consists of red, purplish-red and grey-red conglomerate, siltstone and shale with minor green shales. The conglomerates are poorly to well sorted and contain fragments up to 20 cm in diameter. These fragments consist of granite, felsic volcanic rocks, red and green sandstones and shales in a fine grained matrix dominated by quartz, hematite, calcite, sericite and chlorite. Granite fragments consist of quartz and saussuritized feldspar displaying graphic intergrowths. Felsic volcanic fragments contain microphenocrysts of plagioclase in a partly sericitized cryptocrystalline quartzofeldspathic groundmass. Sandstone clasts are well sorted and are compositionally dominated by subrounded quartz with minor feldspar in a hematitic matrix. Shale fragments are very similar to the red and green shales of the Arisaig Group. The siltstone and shale units that are interlayered with the conglomerates are micaceous. The siltstones commonly display graded bedding and cross-bedding. The shales characteristically contain abundant pale green calcareous nodules up to 5 cm in diameter. Poorly preserved brachiopods were found in the green shales near Lochaber Lake. The assignment of these rocks to the Knoydart Formation is based on the strong lithological comparison of these rocks to the type section at Knoydart (compare Boucot et al., 1974).

At Lochaber Lake, two cleavages, a C cleavage that is parallel to bedding and an S cleavage which is generally at about 30± to bedding (see Berthe et al., 1979) are well developed in all lithologies adjacent to the Lochaber Lake Fault. Fragments are intensely flattened and stretched and are elongate parallel to the S cleavage.

Correlations

The stratigraphy of the Arisaig Group in the type area (Boucot et al., 1974) and in the Antigonish Highlands (this study) is shown in Table 6. In general, the two sequences correlate favourably and similar definitions of the boundaries between formations may be applied in both areas. Selected stratigraphic sections of uppermost Ordovician to Lower Silurian rocks of the Antigonish Highlands are shown in Figure 48. For comparative purposes, an idealized section of the type area (section A) of the Arisaig Group (compiled from Boucot et al., 1974; Keppie et al., 1978; Pickerill and Hurst, 1983; Hurst and Pickerill, 1986) is also presented.

The Bears Brook Formation is considered to be laterally equivalent to the Dunn Point and McGillivray Brook formations. All of these formations contain abundant volcanic rocks and display evidence of subaerial deposition. The most important lithological difference between these formations is that the Bears Brook Formation contains abundant red conglomerates and arkoses and displays significant lateral variations in the relative abundance of each lithology and in thickness (Fig. 48). The Dunn Point and Bears Brook formations both contain bimodal volcanic rocks (see the section on Geochemistry below). The chemistry of the volcanic rocks in these formations is virtually indistinguishable. The geochemistry of the volcanic rocks of the Bears Brook Formation probably represents a continental within-plate extensional tectonic environment. Keppie et al. (1978) propose a similar interpretation for the Dunn Point and McGillivray Brook formations.

The spatial and temporal association of both of these volcanic sequences with the Beechill Cove Formation indicates that these formations document a broadly cogenetic sequence. This is supported by the similarity between the sedimentary rocks of the Bears Brook and the lowermost facies of the Beechill Cove Formation, and by the probable latest Ordovician or Early Silurian age indicated by the Rb-Sr isotopic data.

It is proposed that the Bears Brook, Dunn Point and McGillivray Brook formations should be included in the Arisaig Group. According to article 19 of the North American Stratigraphic Code a revision of the boundary of a unit is justifiable "if a minor change in the boundary or content will make the unit more natural or useful". Although the definition of the Arisaig Group is well established in the literature as a succession of fossiliferous siliciclastic rocks, the revision is proposed because:

1) The contact between the Bears Brook and Beechill Cove formations is conformable in the Antigonish Highlands. This indicates that the disconformity between the Beechill Cove Formation and the underlying volcanic rocks in the Arisaig area represents another example of overstepping due to marine transgression rather than a significant break in the sequence.

2) Rhyolites occur within the Beechill Cove Formation within the Antigonish Highlands. Thus the exclusion of volcanic rocks from the Arisaig Group is not warranted. The rhyolites are chemically distinct from the rhyodacites of the Bears Brook and Dunn Point formations but they are virtually indistinguishable from the rhyolites of the McGillivray Brook Formation. The rhyolites in both formations are anomalously high in Si and low in Mg, Fe, Ti and Sr (see the section on Geochemistry below).

3) The geochemistry of the volcanic rocks records a continental rift-related tectonic setting. A marine transgression at the base of the Silurian documented by the first appearance of marine fossiliferous strata is consistent with this environment. Thus the volcanic and siliciclastic rocks probably form a cogenetic sequence.

It is therefore suggested that the inclusion of the Dunn Point, McGillivray Brook and Bears Brook formations into the Arisaig Group makes the definition of the Arisaig Group "more natural and useful".

For similar reasons, the Knoydart Formation should also be considered part of the Arisaig Group because as noted by Boucot et al. (1974) the Knoydart Formation rests conformably on the Stonehouse Formation. Although the Knoydart Formation is dominated by red beds (in contrast to the underlying Stonehouse Formation), the exclusion of the Knoydart Formation from the Arisaig Group on lithological grounds is not warranted because red beds also occur within the Moydart Formation. It is suggested that the Knoydart Formation forms part of a cogenetic sequence within the Arisaig Group and represents a regressive stage in the depositional history.

In summary, the revised Arisaig Group represents a continuous sequence of uppermost Ordovician to Lower Devonian sediments. It unconformably overlies Precambrian and Cambrian-Lower Ordovician sequences and is unconformably overlain by the Upper Devonian McAras Brook Formation.

In general, the stratigraphy of the Beechill Cove and Ross Brook formations in the Antigonish Highlands compares favourably with the type area. The main difference is the occurrence of a thin volcanic unit (generally felsic) within the Beechill Cove Formation in the Antigonish Highlands (sections F and H, Fig. 48). Two dominant lithologies have been recognized in the Beechill Cove Formation in the Antigonish Highlands: a lower member dominated by red fluviatile conglomerates and arkoses; and an upper member which consists of mottled massive fossiliferous shallow-marine siltstones. Locally (e.g. the Glencoe Brook area), quartzites and laminated siltstones also occur. In the type area, near Arisaig, Pickerill and Hurst (1983) document six facies: facies 1 consists of a thin oligomictic conglomerate deposited as a transgressive beach lag; facies 2 is composed of red shales deposited in a shoreface environment; facies 3 is dominated by mottled mudstones interpreted as a shallow subtidal deposit; facies 4 is a regular layered turbiditic siltstone deposited in a shelf environment with a periodically high energy environment of deposition; facies 5 contains lenticular siltstones deposited in a subtidal environment; and facies 6 is a laminated shale deposited in a low energy environment (see Fig. 2 of Pickerill and Hurst, 1983).

The lower lithology of the Beechill Cove Formation in the Antigonish Highlands is probably a lateral variation of facies 1 and 2 of Pickerill and Hurst (1983). In both the Antigonish Highlands and Arisaig these lithologies overlie volcanic rocks and underlie fossiliferous siltstones. However, in the type area, facies 1 has been interpreted as a transgressive beach lag deposit (Pickerill and Hurst, 1983) whereas the lateral equivalent in the Antigonish Highlands represents a fluviatile environment. The upper lithology in the Antigonish Highlands is very similar to facies 4 in that it is laterally continuous and commonly contains thick graded shell layers at the base of the siltstone units. These layers contain the *Lepostrophia beechhillensis* faunal association (R.K. Pickerill, pers. comm., 1985) that is typical of the Beechill Cove Formation.

Boucot et al. (1974) divided the Ross Brook Formation into three members. The lower member consists of black shales, the middle member is distinguished by a gradual change in colour from black to grey together with interbedded siltstones and the upper member contains an increased proportion of siltstones interbedded with the shales. The lower member of the Ross Brook Formation in the Antigonish Highlands is very similar to the lower and middle members of Boucot et al. (1974), whereas the upper member of the respective areas are virtually identical.

The French River Formation was proposed by Maehl (1961) for the rocks that conformably overlie the Ross Brook Formation on French River in the northwest Antigonish Highlands. On paleontological and lithological grounds

Boucot et al. (1974) defined the French River Formation in the equivalent stratigraphic position in the Arisaig area (i.e. where a sharp colour change separates the lower grey shales of the Ross Brook Formation and the green siltstones of the French River Formation).

The Doctors Brook Formation is not represented in the Antigonish Highlands. This is probably because the exposures in the Antigonish Highlands are not of sufficient quality to identify this formation. Furthermore, the definition of the lower contact of the Doctors Brook Formation with the French River Formation in the Arisaig area is based on the relative dominance of siltstones and shales and this definition may have only local applications. It is doubtful whether such a definition can have regional applications. The French River Formation as defined by Maehl (1961) and accepted here may encompass both the French River and Doctors Brook formations as defined by Boucot et al. (1974).

Boucot et al. (1974) defined the base of the McAdam Formation by the occurrence of black shales. This lithology was not found in the Antigonish Highlands; the lower boundary with the French River Formation is defined on the greater abundance of shales in the McAdam Formation. This makes the contact difficult to define in poorly exposed and intensely deformed areas (e.g. near the Chedabucto Fault).

The definitions of the Moydart, Stonehouse and Knoydart formations as described by Boucot et al. (1974) apply to the Antigonish Highlands. Maehl (1961) and Benson (1967, 1974) have described fauna similar to that of Boucot et al. (1974) in each of the respective formations.

Depositional environment

In general, the stratigraphic sequence records a marine transgression during the lowermost Silurian with subaerially deposited rocks (Bears Brook and lower Beechill Cove formations) overlain by shallow marine clastic rocks (upper Beechill Cove and Ross Brook formations). The presence of welded ignimbrites, abundant red clastic sediments and rapid lateral variations in lithology and in the thickness of the sequence suggest a subaerial environment of deposition for the Bears Brook and lower Beechill Cove formations in which there was substantial paleotopographic control. The textures of the red conglomerates and arkoses suggest a fluviatile environment of deposition. This is consistent with the red colour, the presence of normal and inverse grading, locally irregular bedding, the abundance of channels up to 5 m across and the abundance of "rip-up" clasts of red shale within the conglomerates.

In detail, the Bears Brook Formation is characterized by rapid and locally controlled lateral lithological variations. For example, sections B and C (Fig. 48), taken only 4 km apart, are dominated by coarse conglomerates (with pebbles up to 10 cm) and by fine arkosic sandstones, respectively. Local variations in lithology, texture and degree of angularity of the fragments displayed in the conglomerate suggest that they have a fanglomeratic component. Volcanic rocks within this formation are highly variable in abundance and are most abundant adjacent to major faults such as the Browns Mountain and Chedabucto faults. The mafic volcanic rocks are relatively subordinate or are absent from many sections. Felsic volcanic rocks (flows, ignimbrites and tuffs) are invariably present and are especially dominant in the southwestern highlands (section H, Fig. 48). These pronounced lateral variations in the distribution of the volcanic rocks indicate a significant paleotopographic control.

The most important facies variations occur north and south of the Browns Mountain Fault, which contrast with the lack of lithological variations across the Hollow Fault. To the south of the Browns Mountain Fault, felsic volcanic rocks are dominant (compare sections A, B and C with sections F and G, Fig. 48). This suggests that the southern Antigonish Highlands was a centre of felsic volcanic activity in the latest Ordovician-earliest Silurian and that the Browns Mountain Fault was either tectonically active or had significant topographic expression (or both) at that time. This interpretation is supported by the presence of a concentric magnetic anomaly and radar imagery pattern in the vicinity of the felsic volcanic rocks to the west of Eden Lake in the southern highlands. The patterns are tentatively attributed to a vent complex.

Pickerill and Hurst (1983) interpret the lower Beechill Cove facies north of the Antigonish Highlands to be a transgressive beach lag deposit. In contrast, in the Antigonish Highlands, volcanic rocks are overlain by a thick sequence of arkose and conglomerate. The similarity of the arkose and conglomerate to those in the Bears Brook Formation suggests that they were probably deposited in a similar fluvial environment to the Bears Brook Formation in contrast to the laterally equivalent black lag rocks to the north of the Hollow Fault. These relationships indicate that the shoreline at the base of the Silurian was probably in the vicinity of the present day trace of the Hollow Fault.

The fossiliferous siltstones of the Beechill Cove Formation represent the first marine rocks of the Arisaig Group. Pickerill and Hurst (1983) interpret the "regular layered" siltstones to be deposited by storm-generated turbidity currents in a subtidal marine shelf sequence, and the laminated siltstones to represent a shallow subtidal environment.

The relative homogeneity of the upper Beechill Cove and Ross Brook formations and the presence of shallow marine fossils indicate that the Antigonish Highlands progressively became submerged. This is supported by the distribution of Silurian rocks at the extremities of the highlands and is further indicated by overstep relationships shown in Figures 46 and 48. For example, the Beechill Cove Formation may rest directly on basement (see D, Fig. 48) or the Ross Brook Formation may rest on the Bears Brook Formation (see I, Fig. 48). In some instances, the contacts between formations of the Arisaig Group are locally unconformable. The above discussion supports the interpretation of Pickerill and Hurst (1983) that the Beechill Cove Formation represents a transgressive sequence. The distribution of the clastic rocks at the margins of the Antigonish Highlands suggests that rocks of the Arisaig Group draped the Antigonish Highlands.

The marine transgression at the base of the Silurian has been associated with a world-wide eustatic marine transgression (e.g. Cant, 1980). The synchroneity between rifting (as indicated by the geochemistry of the volcanic rocks, Keppie et al., 1979; Murphy, 1987) and the eustatic rise in sea level may be responsible for the almost complete sequence of uppermost Ordovician to Lower Devonian strata deposited in the Antigonish-Arisaig area. Throughout the Silurian the sequence is dominated by shallow marine siliciclastic rocks. There are no large scale coarsening or fining upward sequences, and fluctuations in the energy of the depositional environment appear to be on a relatively small scale. This suggests that sedimentation was able to keep pace with eustatic changes in sea level. Cant (1980) and Pickerill and Hurst (1983) suggest that sedimentation of the Arisaig Group in the type area was governed by periodic storm activity. A similar model is possible in the Antigonish Highlands.

Uppermost Silurian to Lower Devonian strata show an increasing abundance of red shale and conglomerate which indicates a gradual shallowing of the environment of deposition and a return to subaerial deposition.

Conclusions

The definition of the Arisaig Group is revised to include the uppermost Ordovician to lowermost Silurian Bears Brook Formation, the laterally equivalent Dunn Point and McGillivray Brook formations and the Lower Devonian Knoydart Formation. This revision means that the Arisaig Group occupies a strategic position in the history of the Appalachians in that it records an almost complete sequence of rocks that were deposited between the Taconic and Acadian orogenic events.

It is recommended that the Charcoal, Sunnybrae, Glencoe Brook and Kerrowgare formations defined by Maehl (1961) for the southern Antigonish Highlands be abandoned because rocks assigned to these formations are similar to those of the northern Antigonish Highlands. Thus these names should be replaced by the better known northern equivalents.

The Arisaig Group in the Antigonish Highlands was deposited in a continental to shallow marine, within-plate extensional environment. Rifting resulted in bimodal volcanism and a marine transgression in which continental clastic rocks are overlain and overstepped by marine fossiliferous rocks. The Bears Brook Formation and the lower member of the Beechill Cove Formation were deposited in a fluvial environment in contrast to the laterally equivalent beach lag rocks to the north of the Hollow Fault. This suggests that the shoreline at latest Ordovician-earliest Silurian time was probably close to the present-day trace of the Hollow Fault. The marine transgression was probably due to the synchroneity of local rifting (Murphy, 1987) and the eustatic rise in sea level (Cant, 1980).

The Beechill Cove Formation north and south of the Hollow Fault is lithologically and paleontologically similar. Pickerill and Hurst (1983) interpreted this formation north of the Hollow Fault as a transgressive marine sequence and an analogous interpretation is offered here for these rocks in the Antigonish Highlands. The lack of coarsening or fining upward sequences within the upper Silurian rocks indicates that sedimentation was able to keep pace with eustatic changes in sea level.

The most significant lateral variations were found across the Browns Mountain Fault rather than the Hollow Fault. This suggests that there may be some difficulties in evaluating regional correlations (i.e. they may be difficult to separate from local controls). Assuming that the Bears Brook Formation was once continuous across the Browns Mountain Fault, displacement across the fault is less than 5 km.

In the simplest sense, the similarities between these successions north and south of the Hollow Fault suggest that they were once continuous. This therefore restricts cumulative post-Early Silurian movement on the Hollow Fault to about 40 km in a dextral sense. This may have significant implications on palinspastic reconstructions. Alternatively, if future work on rocks of this age in the Appalachians shows that these types of successions are common, then significantly more movement may be accommodated.

Geochemistry

Analytical notes

A total of 16 mafic volcanic rocks and 22 felsic volcanic rocks of the Bears Brook Formation, and 7 felsic volcanic rocks of the Beechill Cove Formation were analyzed for major and trace elements by X-ray fluorescence at St. Mary's University, Halifax. The precision and accuracy of all elements are about 5 percent each with the exception of Na_2O and MgO which are between 5 and 10 percent (K. Cameron, pers. comm., 1985). A total of 10 mafic volcanic rocks and 4 felsic volcanic rocks (3 from the Bears Brook Formation and 1 from the Beechill Cove Formation) were analyzed for rare earth elements at X-Ray Assay Laboratories, Toronto. The precision and accuracy for the rare earth element data are about 10 percent each.

A summary of the geochemistry of the volcanic rocks is presented in Table 7. The Bears Brook Formation is bimodal with respect to SiO_2, with a gap from 52.8 to 66.0 weight percent (on a volatile-free basis). The generally low SiO_2 and high Fe_2O_3t and MgO contents confirm the basaltic and mafic nature of many of the volcanic rocks, whereas the SiO_2 content in the felsic volcanic rocks classifies them as rhyodacites or rhyolites (see Fig. 52b).

Mafic rocks

The mafic volcanic rocks of the Bears Brook Formation have a highly variable major and trace element geochemistry (Table 7a). For example, Fe_2O_3t ranges from 9.9 to 20.5 weight percent, TiO_2 from 1.5 to 4.3 weight percent and P_2O_5 from 0.2 to 2.6 weight percent. These elements decrease with increasing SiO_2, on a volatile-free basis (Fig. 50). Zr ranges from 60 to 260 ppm and has no apparent correlation with SiO_2 (Fig. 50). The ranges and average concentrations of many

Table 7a. Geochemistry of the mafic volcanic rocks of the Bears Brook Formation

	70131	70200	70262	70212	70214	70215	70232	27114
SiO_2	49.00	42.60	44.70	39.90	42.90	44.50	44.60	46.50
TiO_2	1.35	2.78	2.58	3.93	2.82	2.74	1.76	3.28
Al_2O_3	19.00	16.00	16.30	15.70	16.10	15.90	16.80	14.10
Fe_2O_3	11.70	12.30	12.30	16.50	15.00	14.40	10.30	14.80
FeO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	5.89	7.16	7.89	7.15	6.32	6.77	6.17	5.96
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CaO	2.20	8.38	4.52	4.52	5.49	4.87	6.17	3.12
Na_2O	5.39	2.49	4.04	3.00	3.16	3.52	4.07	3.69
K_2O	0.72	0.38	0.13	0.13	0.67	0.55	1.25	0.12
P_2O_5	0.18	0.47	0.85	1.38	0.49	0.45	55.00	0.69
H_2O	4.08	3.77	5.70	6.16	5.85	4.93	7.00	6.00
CO_2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
TOTAL	99.53	96.33	99.01	98.37	98.80	98.63	98.67	98.26
Rb	10	20	10	10	20	40	30	10
Sr	830	510	430	470	330	460	450	270
Y	20	30	30	40	30	10	10	41
Zr	80	130	200	160	120	110	140	260
Nb	10	30	50	20	30	30	40	32
Ba	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Ni	3	76	90	46	64	73	79	12
Cr	60	180	240	80	180	190	200	20
V	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Ga	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Sc	27	41	34	34	37	37	28	n.d.
Hf	n.d.	4	5	4	4	3	4	n.d.
La	13	18	31	29	17	16	33	n.d.
Ce	30	44	67	70	37	38	63	n.d.
Nd	10	20	30	40	20	20	30	n.d.
Sm	4.3	5.6	7.9	10.5	5.3	4.9	6.4	n.d.
Eu	1.4	2.2	3	4.8	2.1	2.3	1.9	n.d.
Yb	2	2	4	4	2	2	3	n.d.
Lu	0.4	0.4	0.5	0.5	0.3	3.7	0.4	n.d.

	70231	70233	70311	80311	70460	70461	43140	42420
SiO_2	44.70	47.90	50.60	38.80	44.30	48.60	40.10	45.80
TiO_2	2.54	1.79	1.57	1.36	1.77	1.64	3.71	2.42
Al_2O_3	16.60	16.90	18.90	18.50	17.90	16.90	15.90	15.10
Fe_2O_3	11.80	10.50	10.10	9.85	11.70	12.00	19.50	14.10
FeO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	5.91	7.12	4.00	3.50	3.76	4.69	3.68	6.48
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CaO	5.07	3.79	5.74	10.30	8.13	8.92	3.89	7.26
Na_2O	4.90	3.90	3.72	2.67	4.24	3.25	2.76	2.38
K_2O	0.10	0.93	1.15	2.42	0.58	0.67	2.80	1.65
P_2O_5	0.84	0.58	0.50	0.38	0.43	0.41	2.56	0.33
H_2O	5.31	5.39	3.85	11.80	6.77	2.16	3.62	2.62
CO_2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
TOTAL	97.82	98.80	99.13	99.58	99.58	99.24	98.52	98.14
Rb	10	20	30	70	10	20	20	60
Sr	350	450	730	120	460	700	140	500
Y	30	30	30	30	20	50	67	30
Zr	210	150	190	60	150	150	400	150
Nb	30	40	30	10	10	20	46	22
Ba	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Ni	92	75	11	67	22	21	2	44
Cr	250	190	70	180	40	39	10	40
V	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Ga	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Sc	33	27	30	39	40	39	n.d.	n.d.
Hf	5	4	5	3	4	4	n.d.	n.d.
La	25	34	42	33	25	25	n.d.	n.d.
Ce	52	70	74	61	52	52	n.d.	n.d.
Nd	30	20	30	20	20	20	n.d.	n.d.
Sm	7.3	6.6	6.8	6.8	6.0	6.0	n.d.	n.d.
Eu	2.7	1.8	n.d.	2.5	2.2	1.9	n.d.	n.d.
Yb	3.0	3.0	3.0	2.0	3.0	3.0	n.d.	n.d.
Lu	0.5	0.5	0.5	0.3	0.5	0.4	n.d.	n.d.

n.d. = not determined

elements (Table 7a) are similar to those of the time-equivalent Dunn Point Formation (Keppie et al., 1979) to the north of the Hollow Fault.

Felsic rocks

The felsic volcanic rocks of the Bears Brook Formation are relatively homogeneous in chemistry although they contain variable Na_2O and K_2O. These volcanic rocks are characterized by low and variable Rb/Sr ratios. They are peraluminous with molecular $Al_2O_3 > Na_2O+K_2O+CaO$ and display LREE enrichment (Fig. 53b) with high La/Sm. Felsic volcanic rocks of the Beechill Cove Formation are geochemically distinct from those of the Bears Brook Formation. They contain higher SiO_2 and on average higher K_2O and lower Na_2O, Al_2O_3, CaO, P_2O_5, TiO_2, Sr, Nb, Y and LREE (Fig. 51 and 53d; and Table 7c).

Magmatic affinity

Although alteration indicates that discrimination diagrams should be used with caution (especially for the metabasalts), some constraints can be offered by combining geochemistry with lithological association. The high Fe_2O_3t contents (i.e. abundant primary magnetite) and high Fe2O3/MgO ratio are inconsistent with a calc-alkalic origin but do suggest a tholeiitic affinity (e.g. Miyashiro, 1974). The relatively high Zr and P_2O_5 contents (Table 7a) and the low Zr/P_2O_5 ratios are typical of alkalic rocks (Fig. 52a, after Floyd and Winchester, 1975). On the other hand, Zr/TiO_2 and Nb/Y ratios are variable and straddle the alkalic and subalkalic (in this case tholeiitic) boundary (Fig. 52b,c). However, the variable Nb/Y ratios also suggest phlogopite fractionation which is typical of alkalic basalts and a mantle source enriched in K and Rb (e.g. Boettcher and O'Neil, 1980). The humped multielement pattern (see Fig. 55) is also consistent with an alkalic affinity. Thus the geochemistry of the mafic rocks is probably transitional between alkalic and tholeiitic suites. The rare earth element profile (Fig. 53a) which displays LREE enrichment with high La/Sm (3 to 5) and moderate La/Yb (6 to 11) ratios is consistent with this interpretation (Basaltic Volcanism Study Project, 1981).

Felsic rocks are peraluminous. Bears Brook felsic volcanic rocks plot in the rhyodacite field whereas the Beechill Cove volcanic rocks are rhyolites (Fig. 52b).

Tectonic setting

The tectonic setting of these rocks is probably within-plate, continental and extensional. For the mafic volcanic rocks, a within-plate origin is suggested by the high concentration of high field strength elements (Fig. 55a), in particular the high TiO_2 and Zr contents (Fig. 54a), and by the high Zr/Y ratios (between 4 and 11, Fig. 54b, after Pearce and Norry, 1979) in the mafic volcanic rocks. The multielement profiles of the felsic volcanic rocks of the Bears Brook and Beechill formations (Fig. 55b,c) show moderate enrichment in LIL elements and slight depletion in high field strength elements relative to ORG. This pattern is typical of attenuated continental crust (Pearce et al., 1984a) and is therefore consistent with a continental rift setting. A continental environment is indicated by the stratigraphy which consists of subaerial to shallow marine clastic rocks. The bimodality of the volcanism with respect to SiO_2 is also consistent with this origin (e.g. Martin and Piwinski, 1972) and indicates an extensional tectonic environment, as does the temporal association of the volcanism and the marine transgression. The relationship

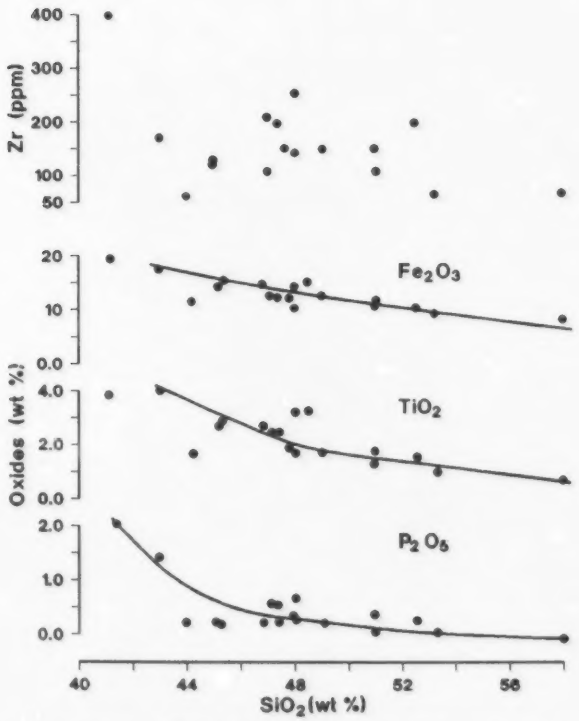


Figure 50. The relationships of Zr, Fe_2O_3t, TiO_2 and P_2O_5 to SiO_2 in the mafic volcanic rocks of the Bears Brook Formation.

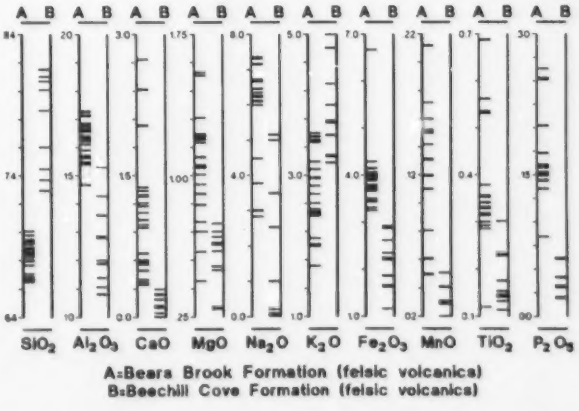


Figure 51. Geochemical comparison between the felsic volcanic rocks of the Bears Brook and Beechill Cove formations. Values are in weight percent.

between facies variations, centres of volcanic activity and major faults suggests that these faults were at least moderately active at this time.

Petrogenesis

Mafic rocks

In detail, much of the variation cannot easily be attributed to primary magmatic trends but is more likely due to alteration as described above. This alteration precludes detailed petrological modelling. In particular Na, K and Rb show a considerable variation (Table 7) that is probably due to alteration.

However, Fe_2O_3t variations and to an extent TiO_2 and P_2O_5 variations may have a magmatic component, as they correlate with the modal abundance of magnetite and apatite. The antipathetic relationship between these elements and SiO_2 (Fig. 50) indicates that their abundances may be controlled by fractionation of magnetite and apatite.

The high Fe_2O_3/MgO ratios suggest that fractionation of ferromagnesian phases such as olivine and pyroxene occurred. Olivine fractionation is also indicated by the relatively low Ni contents, and augite fractionation is indicated by the wide range in Cr contents and by the presence of augite microphenocrysts. The LREE of the mafic volcanic rocks (Fig. 53a) is consistent with partial melting or fractional crystallization with respect to garnet in a garnet lherzolite source. Alternatively, the rare earth element profiles could be attributed to partial melting of a metasomatized mantle. For example, the wide range in Zr/Y and the low Zr/Nb ratios (Fig. 56) indicate that the basalts may have been derived from metasomatized mantle typical of the source of alkalic basalts (e.g. Boettcher and O'Neil, 1980; Le Roex et al., 1983).

Felsic rocks

The Bears Brook rhyodacites are interlayered with the basalts apparently implying a cogenetic relationship. The two models most commonly invoked to explain the origin of felsic

Table 7b. Geochemistry of the felsic volcanic rocks of the Bears Brook Formation

	410-3A	G02-12B	G02-M	G03-5	31A-9	41-TC1B	32-2K	64-1G1	27A-5	84-4	84-5
SiO_2	67.7	64.7	64.2	64.0	68.4	68.7	65.5	60.0	68.3	69.5	69.3
TiO_2	0.37	0.68	0.63	0.63	0.29	0.28	0.56	1.08	0.30	0.32	0.34
Al_2O_3	15.7	16.2	16.2	15.9	15.6	15.5	14.7	16.0	15.4	15.8	16.3
Fe_2O_3'	4.04	4.13	4.00	3.94	3.36	3.25	6.57	7.39	3.47	3.47	3.69
MgO	1.02	1.21	1.22	1.30	0.91	1.05	0.70	2.40	1.23	1.06	1.20
MnO	0.12	0.12	0.15	0.11	0.17	0.15	0.16	0.20	0.11	0.13	0.15
CaO	0.68	2.03	2.42	2.73	0.99	1.13	1.22	2.82	0.96	1.36	0.61
Na_2O	6.68	3.80	3.03	2.82	6.52	6.27	4.56	4.29	6.26	6.64	7.20
K_2O	2.17	3.20	3.43	3.52	2.44	2.60	3.46	2.92	2.06	1.72	2.02
P_2O_5	0.20	0.26	0.25	0.25	0.13	0.13	0.08	0.38	0.14	0.15	0.16
BaO	n.d.	n.d.	n.d.	n.d.	0.12	0.12	0.17	0.11	0.08	n.d.	n.d.
LOI	1.39	3.77	4.16	4.47	1.20	1.28	2.00	2.16	1.39	2.08	1.15
TOTAL	100.07	100.10	99.69	99.67	100.13	100.46	99.68	99.75	99.70	102.19	102.10
Rb	50	120	105	100	70	10	60	70	50	35	36
Sr	106	220	210	104	370	60	140	310	460	272	321
Y	18	20	18	30	23	24	117	51	23	21	22
Zr	176	230	230	220	160	470	730	260	190	172	172
Nb	20	30	30	25	5	6	92	11	9	8	9
Ba	502	n.d.	583	754	n.d.	n.d.	n.d.	n.d.	900	478	784
Ni	n.d.	n.d.	1	n.d.	<2	<2	2	2	<2	1	3
Cr	39	50	32	40	10	<2	10	10	10	22	28
V	17	51	51	48	n.d.	n.d.	n.d.	n.d.	n.d.	20	21
Ga	15	16	16	16	n.d.	n.d.	n.d.	n.d.	n.d.	13	14
Sc	5.6	6.9	7.0	6.6	n.d.	n.d.	n.d.	n.d.	5.0	4.1	n.d.
Hf	5	6	6	7	n.d.	n.d.	n.d.	n.d.	5	4	n.d.
Th	14	11	12	n.d.	n.d.	n.d.	n.d.	n.d.	4.9	3	4
La	35.6	42.6	42.4	45.5	n.d.	n.d.	n.d.	n.d.	30.7	30.9	n.d.
Ce	55	73	74	80	n.d.	n.d.	n.d.	n.d.	56	55	n.d.
Nd	30	30	30	20	n.d.	n.d.	n.d.	n.d.	20	10	n.d.
Sm	5.5	5.6	5.6	5.9	n.d.	n.d.	n.d.	n.d.	4.4	4.2	n.d.
Eu	1.4	1.3	1.3	1.8	n.d.	n.d.	n.d.	n.d.	1.2	1.2	n.d.
Yb	3	3	3	3	n.d.	n.d.	n.d.	n.d.	2	2	n.d.
Lu	0.49	0.51	0.51	0.51	n.d.	n.d.	n.d.	n.d.	0.42	0.42	n.d.

n.d. = not determined

volcanic rocks displaying this relationship are fractional crystallization of the basaltic magma and partial melting of the crustal rocks. The data presented here cannot unequivocally distinguish which of these two models is the correct one. The felsic volcanic rocks contain lower MgO, Fe_2O_3t, CaO, P_2O_5 and TiO_2 contents than the mafic volcanic rocks. If these rocks were related by crystal fractionation, then this would demand efficient fractionation of phases such as pyroxenes, magnetite and apatite. This should lead to a significant enrichment in the incompatible elements such as Zr, Nb and Rb. However, mafic and felsic volcanic rocks show a considerable degree of overlap in the concentrations of these elements (Table 7). Moreover, the higher Y values of the rhyodacites is inconsistent with apatite fractionation. In addition, if the relative surface area abundance of the volcanic rocks approximates the relative volumes of felsic and mafic magmas, then crystal fractionation cannot directly account for the dominance of felsic volcanic rocks. The above arguments indicate that although the Bears Brook volcanic rocks are cogenetic, they are not comagmatic. Facies relationships (Fig. 48), in which mafic volcanic rocks are thicker in the north whereas felsic volcanic rocks are thicker in the south, suggest different source areas and are inconsistent with a comagmatic model.

The most likely origin of the rhyodacites is crustal anatexis. This model accounts for the facies relationships and the relative volumes of mafic and felsic volcanic rocks in the area. Geochemical evidence supporting this origin is difficult to obtain since the source area is not exposed. Rb-Sr isotopic analyses of the samples display too much scatter for reliable estimation of a $Sr^{87}/^{86}$ initial ratio (R.F. Cormier, pers. comm., 1986). However, the rhyodacites are similar in composition to felsic volcanic and intrusive rocks that range in age from Precambrian to Devonian in the Antigonish Highlands. These felsic suites are all characterized by relatively low concentrations of incompatible elements such as Rb, Zr and Nb and by relatively high Sr (see for example, Table 7c), and their genesis has also been attributed to crustal anatexis (see the section Precambrian Rocks-Geochemistry).

Table 7b. (cont.)

	84-6	84-7	84-9	84-10	84-11	84-12	84-13	84-14	84-15	84-16	84-17
SiO_2	68.7	68.1	67.2	68.0	67.6	68.9	68.5	70.1	69.8	69.3	68.5
TiO_2	0.33	0.33	0.35	0.35	0.33	0.34	0.35	0.33	0.33	0.32	0.32
Al_2O_3	16.6	16.3	17.3	16.9	16.1	17.1	16.8	16.8	16.6	16.6	16.4
Fe_2O_3[1]	3.68	3.77	4.12	4.02	3.90	3.81	3.89	3.48	3.58	3.76	3.45
MgO	1.54	1.09	0.70	0.85	1.17	0.76	1.53	1.18	1.05	0.60	0.96
MnO	0.13	0.14	0.08	0.06	0.16	0.05	0.21	0.15	0.16	0.06	0.13
CaO	0.42	1.19	1.38	0.57	1.30	0.36	0.60	0.38	0.61	0.51	1.04
Na_2O	6.00	6.46	6.22	6.10	6.15	6.26	6.14	7.38	7.38	6.71	6.45
K_2O	2.80	2.50	2.86	3.49	2.57	3.64	2.89	2.40	2.46	2.89	2.69
P_2O_5	0.16	0.16	0.17	0.16	0.16	0.15	0.17	0.16	0.15	0.16	0.15
BaO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
LOI	1.62	1.38	1.69	1.31	1.31	1.38	1.46	0.92	1.08	0.92	1.38
TOTAL	101.99	101.35	101.99	101.72	100.81	102.73	102.53	103.20	103.13	101.85	101.47
Rb	62	50	73	94	49	101	56	45	41	67	58
Sr	325	325	298	304	337	283	290	244	307	305	360
Y	22	23	27	24	22	22	23	22	23	25	23
Zr	173	174	186	184	170	180	165	186	183	185	185
Nb	8	9	10	9	9	9	8	9	9	10	9
Ba	930	1007	786	1080	950	1054	988	752	928	947	975
Ni	7	1	4	3	5	0	2	4	2	2	3
Cr	21	19	19	21	26	14	21	23	28	22	18
V	18	18	11	13	17	15	22	15	14	17	11
Ga	14	14	14	16	15	15	14	15	14	15	15
Sc	n.d.	n.d.	n.d.	4.6	n.d.	4.5	n.d.	n.d.	n.d.	n.d.	n.d.
Hf	n.d.	n.d.	n.d.	5	n.d.	4	n.d.	n.d.	n.d.	n.d.	n.d.
Th	2	5	5	3	5	5	5	7	5	5	4
La	n.d.	n.d.	n.d.	34.0	n.d.	28.1	n.d.	n.d.	n.d.	n.d.	n.d.
Ce	n.d.	n.d.	n.d.	59	n.d.	48	n.d.	n.d.	n.d.	n.d.	n.d.
Nd	n.d.	n.d.	n.d.	20	n.d.	20	n.d.	n.d.	n.d.	n.d.	n.d.
Sm	n.d.	n.d.	n.d.	4.8	n.d.	3.8	n.d.	n.d.	n.d.	n.d.	n.d.
Eu	n.d.	n.d.	n.d.	1.7	n.d.	3.1	n.d.	n.d.	n.d.	n.d.	n.d.
Yb	n.d.	n.d.	n.d.	3	n.d.	3	n.d.	n.d.	n.d.	n.d.	n.d.
Lu	n.d.	n.d.	n.d.	0.41	n.d.	0.41	n.d.	n.d.	n.d.	n.d.	n.d.

n.d. = not determined

Table 7c. Geochemistry of the felsic volcanic rocks of the Beechill Cove Formation

	G02-19	G03-12	G03-13	G03-14	38-1	57-88	62-20
SiO_2	78.0	78.4	78.1	75.1	79.9	71.7	78.7
TiO_2	0.2	0.1	0.1	0.2	0.1	0.3	0.1
Al_2O_3	13.60	11.40	11.90	12.90	10.80	15.30	12.00
Fe_2O_3	1.85	2.55	2.41	2.78	1.57	2.85	1.16
MgO	0.71	0.50	0.52	0.74	0.64	0.67	0.64
MnO	0.03	0.03	0.02	0.04	0.03	0.03	0.04
CaO	0.24	0.19	0.13	0.30	0.05	0.10	0.26
Na_2O	0.16	0.08	0.06	0.13	0.25	3.48	2.55
K_2O	4.30	3.30	3.56	3.98	4.78	3.87	3.38
P_2O_5	0.02	0.03	0.03	0.04	0.03	0.06	0.02
LOI	2.70	2.47	2.70	2.77	1.84	1.84	1.80
TOTAL	99.6	99.1	99.6	99.0	100.0	100.1	100.5
Rb	120	100	110	120	140	120	130
Sr	15	12	110	110	10	60	20
Y	38	12	20	40	27	58	28
Zr	120	114	110	110	110	280	90
Nb	28	30	20	10	5	16	10
Ba	731	525	n.d.	n.d.	n.d.	n.d.	n.d.
Ni	4	5	n.d.	n.d.	<2	<2	4
Cr	42	45	60	60	10	10	10
V	1	13	n.d.	n.d.	n.d.	n.d.	n.d.
Ga	14	14	n.d.	n.d.	n.d.	n.d.	n.d.
Sb	n.d.	n.d.	n.d.	n.d.	n.d.	13	4.8
La	n.d.	n.d.	n.d.	n.d.	n.d.	48.8	33.2
Ce	n.d.	n.d.	n.d.	n.d.	n.d.	99	58
Nd	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Sm	n.d.	n.d.	n.d.	n.d.	n.d.	8.4	4
Eu	n.d.	n.d.	n.d.	n.d.	n.d.	1.8	0.6
Yb	n.d.	n.d.	n.d.	n.d.	n.d.	5	3
Lu	n.d.	n.d.	n.d.	n.d.	n.d.	0.9	0.45
Hf	n.d.	n.d.	n.d.	n.d.	n.d.	9	3

n.d. = not determined

The Beechill Cove rhyolitic volcanic rocks have a very distinct geochemistry (Fig. 51). Arguments similar to those above indicate that they are unlikely to be comagmatic with the mafic magma. Of major importance is whether they were derived from the rhyodacites of the Bears Brook Formation or whether they represent a separate magmatic event.

There is much discussion in the literature about the derivation of SiO_2-rich rhyolites from rhyodacites by thermogravitational diffusion (e.g. Hildreth, 1981) where chemical transport is facilitated by the transfer of volatiles (e.g. halogens and CO_2) to the top of the magma chamber. Transport by the halogens should enrich the magma in Zr, Nb, Y and REE (Hildreth, 1981) and transport by CO_2 should enrich the melt in LREE. However, the opposite trends are observed; these elements are depleted in the rhyolites.

Alternatively, if they are related to the rhyodacites by crystal fractionation, then CaO, Al_2O_3, TiO_2, P_2O_5, Sr and possibly REE would have to be extremely efficiently removed (Fig. 51 and 53b; Table 7). Fractionation of plagioclase and a ferromagnesian phase (possibly hornblende) would agree with petrographical data, and together with the fractionation of monazite ± zircon, would account for many of these trends. Fractionation of a rhyodacitic magma adjacent to the walls of a cooling magma chamber

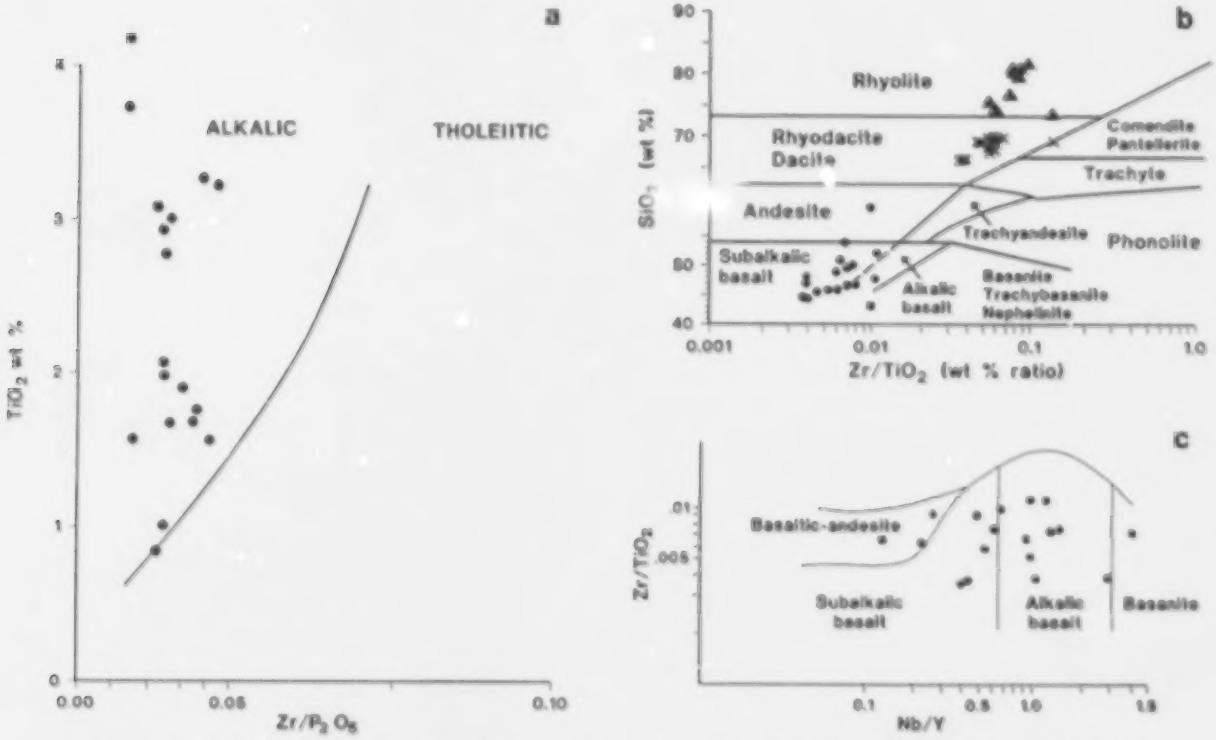


Figure 52. Magmatic affinity of the volcanic rocks of the Bears Brook and Beechill Cove formations (after Floyd and Winchester, 1975; Winchester and Floyd, 1977). (**a**) TiO_2 versus Zr/P_2O_5 (after Floyd and Winchester, 1975), (**b**) SiO_2 versus Zr/TiO_2 and (**c**) Zr/TiO_2 versus Nb/Y. Filled circles-mafic volcanic rocks, Bears Brook Formation, crosses-felsic volcanic rocks, Bears Brook Formation, triangles-Beechill Cove Formation. ALK=alkalic basalts, THOL=tholeiitic basalts.

followed by accumulation of the less dense, fractionated magma at the roof of the chamber is the model suggested for the genesis of the rhyolites.

Summary and discussion

The Arisaig Group in the Antigonish Highlands was deposited in a continental within-plate extensional environment. This resulted in bimodal volcanism and a marine transgression in which continental clastic rocks are overlain and overstepped by marine fossiliferous rocks. Two geochemically distinct felsic volcanic units can be distinguished with rhyodacites overlain by rhyolites. These findings are generally consistent with studies of the Arisaig Group to the north of the Hollow Fault. Keppie et al. (1978) interpreted the uppermost Ordovician-lowermost Silurian volcanic rocks as being within-plate and rift related. The geochemistry of these volcanic rocks is remarkably similar to that of the Antigonish Highlands; they contain interlayered alkalic-tholeiitic mafic and rhyodacitic felsic rocks overlain by an SiO_2-rich rhyolite. From the available data it appears that the rhyodacites were derived by crustal anatexis and that the rhyolites were magmatically derived from the rhyodacites.

The Beechill Cove Formation north and south of the Hollow Fault is lithologically and paleontologically similar. Pickerill and Hurst (1983) interpreted this formation north of the Hollow Fault as a transgressive sequence, and a similar interpretation is proposed here for these rocks in the Antigonish Highlands south of the Hollow Fault.

In the simplest sense, the similarities between these successions north and south of the Hollow Fault suggest that they were once continuous. This therefore restricts cumulative post-Early Silurian movement on the Hollow Fault to about 40 km. This may have significant implications on palinspastic reconstructions. Alternatively, if future work on rocks of this age in the Appalachians shows that these types of successions are common, then significantly more movement may be accommodated.

In conclusion, there is no evidence in either the stratigraphic or magmatic record of a compressive regime in these latest Ordovician-earliest Silurian rocks in the Antigonish Highlands. Rather the environment is one of extension. At this stage it is premature to evaluate whether the continental extensional environment proposed here may represent a localized event or whether analogues may be found across the

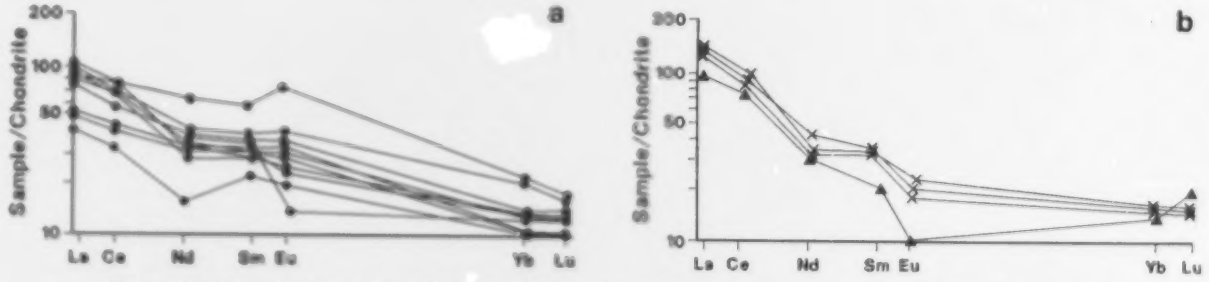


Figure 53. Chondrite-normalized REE patterns for (a) mafic volcanic rocks of the Bears Brook Formation and (b) felsic volcanic rocks of the Bears Brook and Beechill Cove formations. Symbols as in Figure 52.

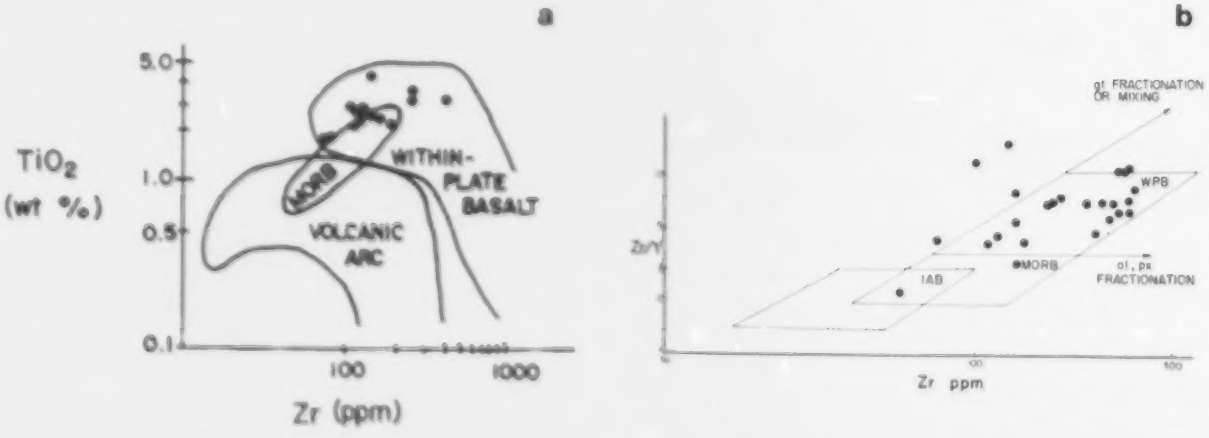


Figure 54. (a) TiO_2 versus Zr diagram (after Pearce et al., 1981) and (b) Zr/Y versus Zr (after Pearce and Norry, 1979) for the mafic volcanic rocks of the Bears Brook Formation. Symbols as in Figure 52. Abbreviations as in Figure 20.

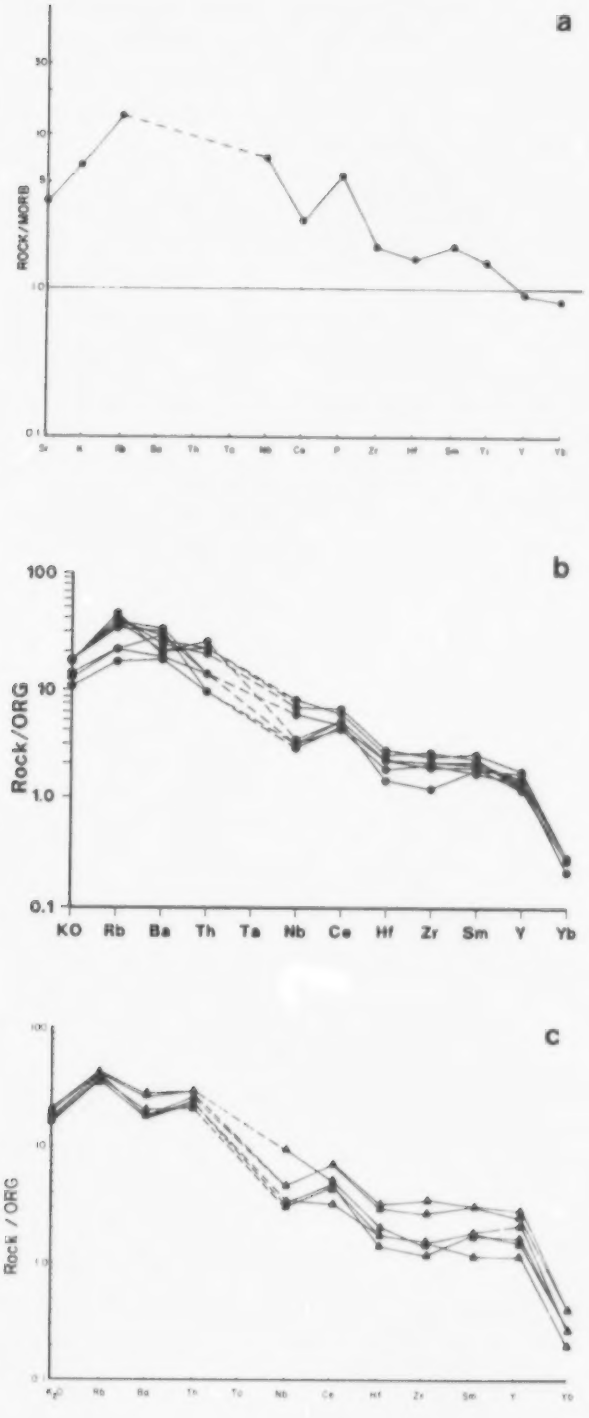


Figure 55. (**a**) Multielement MORB-normalized diagram for the average Bears Brook mafic volcanic rocks; ORG-normalized multielement diagrams for felsic volcanic rocks of (**b**) Bears Brook and (**c**) Beechill Cove formations.

orogen either as a number of local pull-apart basins or as a more regional continental rift. The distinctive geochemical signatures of the volcanic rocks in the Antigonish Highlands may provide constraints for future studies. Recent work on Upper Ordovician-Lower Silurian rocks elsewhere in the Avalon Terrane may shed some light on the regional tectonic significance of these rocks. On the basis of paleomagnetic data (Briden et al., 1988), Keppie (1989) suggested Late Ordovician-Early Silurian time for sinistral accretion of the Avalon Terrane to North America. This interpretation is supported by recent geochronological and kinematic data from the Early Silurian Kingston Dyke Complex in southern New Brunswick (Doig et al., 1990). Thus the geochemistry of the Arisaig Group may represent transtensional conditions associated with the accretion of the Avalon Terrane to North America.

Deformation

Folds

Uppermost Ordovician-Lower Devonian rocks of the Arisaig Group are only moderately deformed in the study area (except when adjacent to faults). In general, these rocks are poorly cleaved, though some of the siltstones and mudstones exhibit a bedding-parallel fissility. These rocks are distributed at the extremities of the Antigonish Highlands (Fig. 46). In areas where they are in depositional contact with Precambrian rocks, they form moderately dipping monoclinal sequences that consistently young away from the central Precambrian core of the highlands (Map 1749A). The Arisaig Group in the Marshy Hope area is deformed into an east-northeast-trending upright shallowly and westerly plunging open fold (Fig. 57). Other than the Marshy Hope area, there is no compelling evidence for regional folding in the area (see the lower hemisphere equal area projections for the northern, western and southern highlands, Fig. 58, 59, 60).

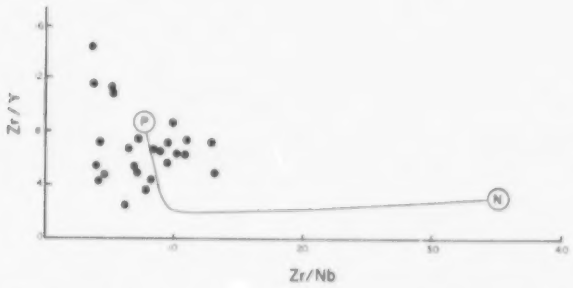


Figure 56. Zr/Y versus Zr/Nb for the mafic volcanic rocks of the Bears Brook Formation. Symbols as in Figure 52. P-basalts derived from an enriched mantle, N-basalts derived from a normal mantle. P-N line is the mixing curve (see Le Roex et al., 1983).

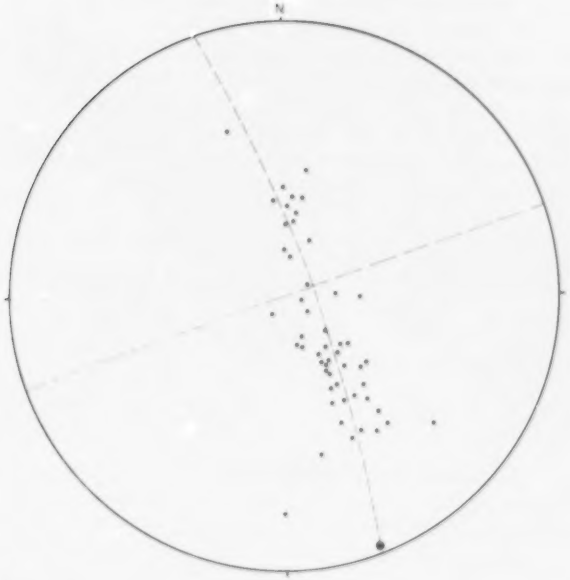


Figure 57. Structural data from the Arisaig Group in the Marshy Hope area.

Figure 59. Structural data from the Arisaig Group in the western Antigonish Highlands.

Figure 58. Structural data from the Arisaig Group in the northern Antigonish Highlands.

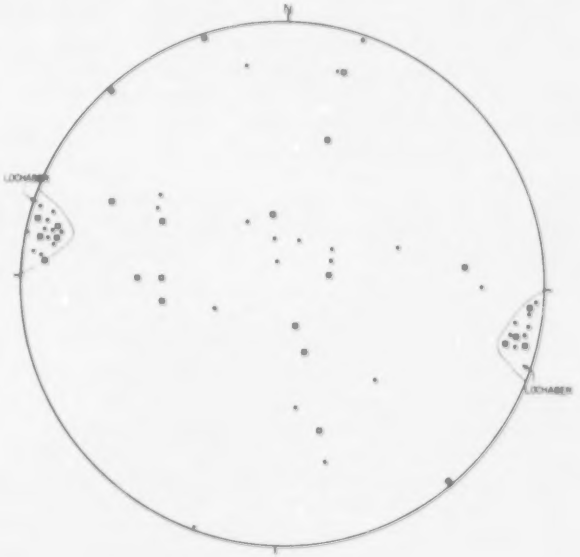


Figure 60. Structural data from the Arisaig Group in the southern Antigonish Highlands.

Synthesis

In comparison to the study area, the type area for the Arisaig Group north of the Hollow Fault is strongly deformed (Boucot et al., 1974). The dominant structure is a northeast-trending moderately southwest-plunging upright to southeast-verging synclinal structure with an axial planar penetrative fabric. The age of these structures is probably Middle Devonian since these folds are not found in Upper Devonian rocks. Similar structures observed in the Cambro-Ordovican rocks (i.e. the F_3 fold in the section on Cambro-Ordovician Rocks--Structure and Metamorphism) and the swing in the trend of F_2 folds in the Georgeville Group (see the section on Precambrian Rocks-Structure and Metamorphism) may also be attributed to this event.

Faults

Faults are of variable ages in the Antigonish Highlands and many may have suffered repeated motions. They may be divided into four categories (Fig. 3): (i) the major bounding faults of the highlands (i.e. the Hollow and Chedabucto faults); (ii) the faults that parallel the bounding faults within the highlands (i.e. the Browns Mountain and Marshy Hope faults); (iii) near-vertical north- to northwest-trending faults at a high angle to the bounding faults (e.g. the Lochaber Lake and Moose River faults); and (iv) thrusts spatially associated with major faults.

i) The Antigonish Highlands are bounded by major transcurrent faults (Fig. 3). The northwestern margin is defined by the northeast-trending Hollow Fault. These faults are considered to be of major regional tectonic significance and have a long history of repeated movements from Precambrian to post-Carboniferous (e.g. Keppie, 1983). The highlands occur between these major lineaments and are affected by brittle and ductile shear zones which are inferred to be a consequence of movement on the major lineaments. The area is consequently dissected by many northeast-trending faults mainly in the northern highlands, and east-trending faults in the southern highlands. Most of these latter faults show limited displacements.

It has been argued in earlier sections that the northeast-trending Hollow Fault may have been repeatedly active since the late Precambrian. The spatial association of the late Precambrian appinitic intrusive rocks with these bounding faults and the lack of appinites elsewhere in the Antigonish Highlands (see section on Economic Geology below) suggest that the appinites may have used these faults as conduits, implying that both the Hollow and Chedabucto faults may have been active at that time. These faults compose part of the Minas Geofracture Fault System which probably represents a zone of repeated strike-slip and dip-slip motions during the Paleozoic (Keppie, 1983). The Hollow and Chedabucto faults were probably active synchronously. These are the most important faults in the area, as they are the bounding faults of the highlands and probably represent major crustal boundaries.

The lithological, stratigraphic and geochemical similarities of Upper Ordovician-Lower Devonian rocks in the Antigonish Highlands with those of the type area of the Arisaig Group suggest that these rocks probably represent an overstep sequence in the tectonic and stratigraphic sense across the Hollow Fault. In the simplest sense, this limits cumulative post-Early Devonian displacment to about 40 km in a dextral sense. However, if similar sequences of this age are found elsewhere in the Appalachians, then significantly more movement may be accommodated.

ii) In the northern part of the map area, a subsidiary set of northeast-trending faults (i.e. parallel to the Hollow Fault) have been identified. The most important of these are the Greendale and Browns Mountain faults. Although it has been argued that these faults were active in the late Precambrian, they are also in fault contact with Carboniferous rocks giving a maximum age for latest displacement (see Boehner and Giles, 1982).

The Browns Mountain Fault in part may represent a major crustal divide within the Antigonish Highlands. It defines both an abrupt change in the style of Precambrian deformation from polyphase deformation in the north to supratenuous folding in the south, and a change in the style of plutonism from felsic-dominated in the south to mafic-dominated in the north. Faults are more numerous in the southern part of the map area, a fact that is presumably related to the brittle nature of the basement rocks. These are mainly vertical faults. Faults adjacent to the Chedabucto Fault tend to be east-northeast- or west-northwest-trending and are generally truncated by the Chedabucto Fault. Many of these can be mapped in the Devono-Carboniferous rocks giving a maximum age of latest displacement.

iii) A set of approximately northeast- and northwest-trending faults are also common between the Browns Mountain and Chedabucto faults. These faults are probably approximately contemporaneous because they truncate rocks of the Arisaig Group and are apparently offset by northeast-trending faults which truncate Carboniferous rocks. Thus they are probably Middle Devonian to Carboniferous in age. They are most strongly defined adjacent to the Chedabucto Fault and apparently die out northward. These faults are characterized by intense shearing, imparting local C and S fabrics to the rocks (see Berthe et al., 1979). Both fabrics are vertically dipping and generally at angles of 30° or less. One of the best examples of this type of fault is the Lochaber Lake Fault which is exposed on the island in Lochaber Lake. The C fabric is defined by planes of dislocation resulting in infinitesimal displacements on crosscutting markers, whereas the S fabric is defined by flattening of pre-existing minerals and by recrystallization of sericite and chlorite. For example, in the Lochaber Lake Fault, the S fabric in the Knoydart Formation is defined in the field by deformed pebbles and calcareous nodules and in thin section by flattening of quartz and feldspar grains and by recrystallization. A mineral lineation associated with this fault is steeply plunging, suggesting that the movement was largely dip-slip.

Despite the intensity of the fabric, these faults have limited displacement. For example, in the Eden Lake area, the Moose River and Garden River faults truncate the Eden Lake Complex with limited offset. The horizontal displacements are apparently limited because the major faults (types (i) and (ii) above) truncate these faults and may have been active at the same time.

The geometry of the C and S fabrics, offsets on quartz veins and fractures and asymmetry of folds were used to obtain the sense of latest offset on these faults. For example, the Lochaber Lake Fault is a reverse fault (east side up). The spatial association of these faults to the Chedabucto Fault suggests that the fault system represents secondary faults due to motion on the Chedabucto Fault. The probable age of displacement (e.g. Middle Devonian to Carboniferous) is

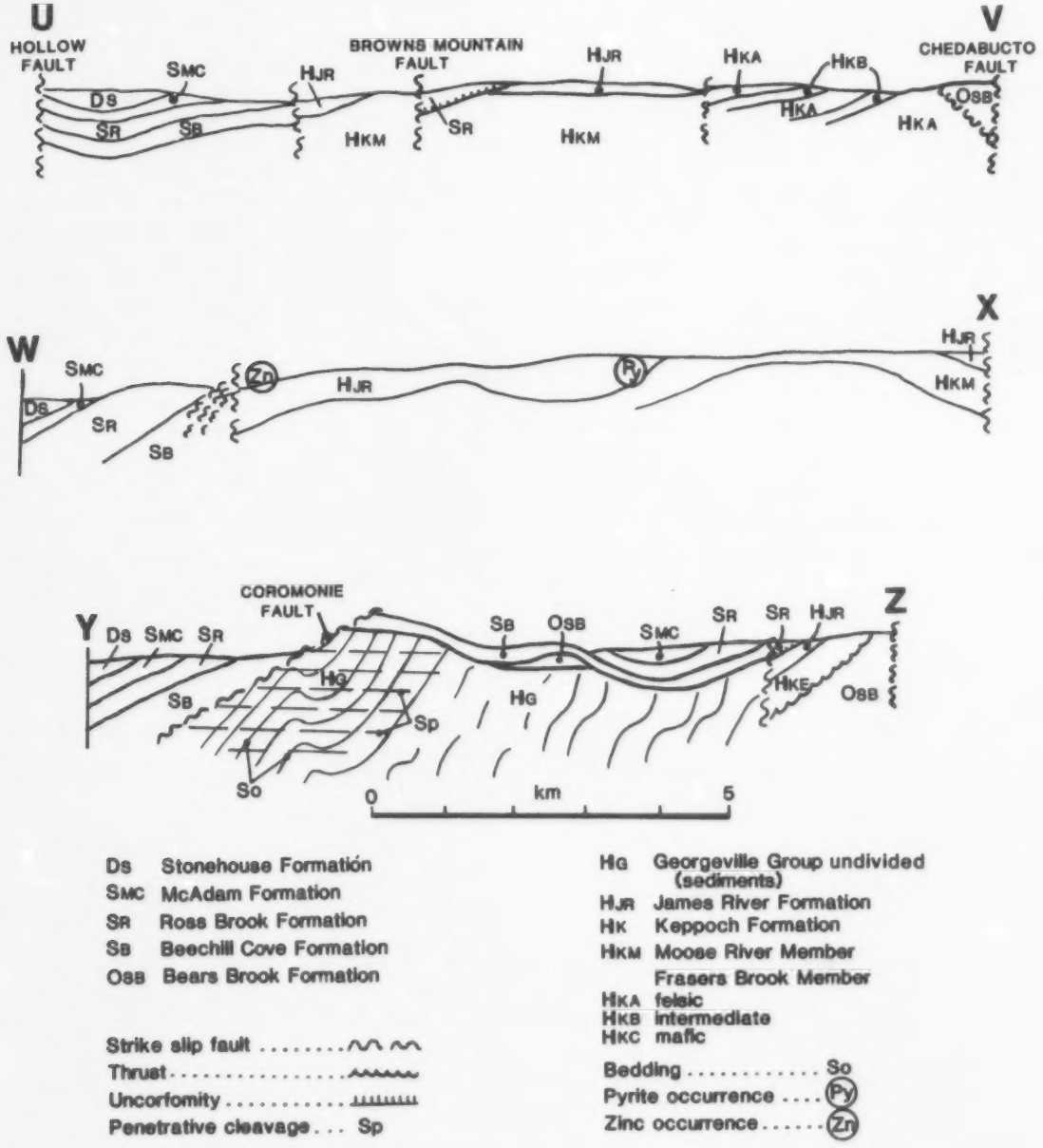


Figure 61. Cross sections in the western Antigonish Highlands. Locations of cross sections given in Figure 46.

roughly synchronous with the probable age of docking of the Meguma Terrane associated with strike-slip movement along the Chedabucto Fault (i.e. Keppie, 1983).

iv) In the western highlands, many of the faults adjacent to the Hollow, Chedabucto and Browns Mountain faults (e.g. the Corromonie Thrust) are probably thrusts (Fig. 3). These faults generally thrust Siluro-Devonian rocks onto Precambrian greywackes, slates and volcanic rocks. For example, immediately to the west of the Corromonie Fault, an intense shallow to moderate west-dipping cleavage occurs in Siluro-Devonian rocks. In the Precambrian rocks east of the fault, cleavage is generally subhorizontal and cuts across folds in these rocks. This suggests that this cleavage (and therefore the thrusting) postdates earlier folding of Precambrian rocks (e.g. Fig. 61).

Imbricate northwest-dipping thrusts and reverse faults associated with the Corromonie Thrust occur on McLellans Brook (Fig. 62). The age of deformation is uncertain, but the youngest unit involved in thrusting in this area is the Lower Devonian Knoydart Formation. Thrusting of Westphalian age associated with movement along the Chedabucto Fault has recently been documented in southern New Brunswick (Nance, 1986). However, the Chedabucto Fault may have suffered continuous movement in the Devono-Carboniferous (e.g. Keppie, 1983) and it is not certain, at this stage, whether these deformations are contemporaneous.

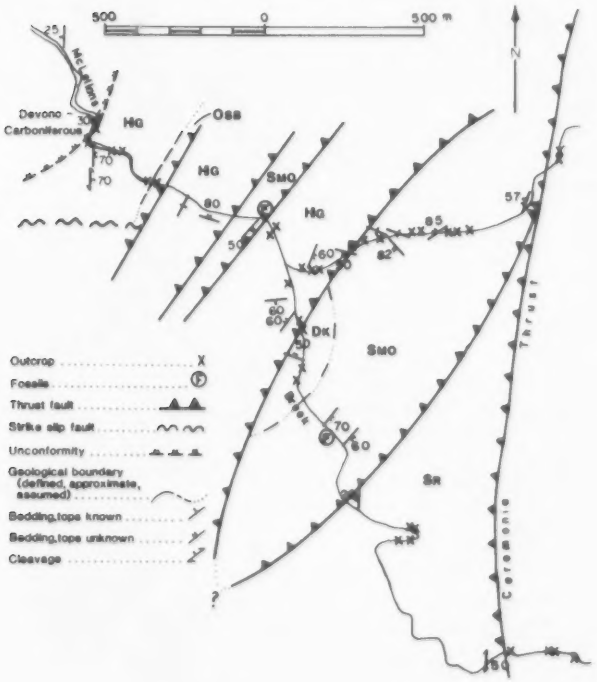


Figure 62. Geological map of the McLellans Brook area. Legend as in Figure 61.

A subhorizontal locally penetrative cleavage is associated with these faults. No exotic stratigraphy occurs and the thrusts probably have only local significance. The faults and the associated cleavage generally dip toward an adjacent major fault. This geometry may indicate the presence of a positive flower structure (see Harding and Lowell, 1979).

Conclusions

Upper Ordovician to Lower Devonian rocks in the Antigonish Highlands consist of bimodal volcanic rocks and red beds at the base overlain by a thick sequence of fossiliferous marine siliciclastic rocks. These rocks belong to the Arisaig Group and correlate well with coeval rocks to the north of the Antigonish Highlands. The definition of the Arisaig Group is revised to include the uppermost Ordovician-lowermost Silurian Bears Brook Formation which occurs at the base, and the Lower Devonian Knoydart Formation which occurs at the top of the succession.

The sequence as a whole is probably related to intra-continental rifting event and may be genetically related to the accretion of the Avalon Terrane to North America. The Bears Brook Formation consists of continental red beds and bimodal volcanic rocks. The volcanic rocks were deposited in a within-plate continental extensional environment. The basalts display alkalic and tholeiitic affinities and the rhyodacites were probably formed by anatexis of the crust. The origin of the younger rhyolites is not clear: they are compositionally distinct from the rhyodacites but may be related to them by late stage differentiation. A marine transgression at the base of the Silurian is indicated by the first appearance of marine fossiliferous rocks and by the overstep relationships exhibited by Silurian strata onto Precambrian basement. The marine transgression may have been associated with a world-wide eustatic rise in sea level and with local rifting. The Knoydart Formation is dominated by fluviatile red beds, indicating a return to subaerial conditions. The distribution of the Arisaig Group rocks around the southern and western margins of the highlands together with the overstep relationships indicate that the Antigonish Highlands became progressively submerged during the Silurian. This was followed by a return to subaerial conditions in the Early Devonian.

Deformation of these rocks by upright shallowly plunging northeast-southwest to east-west folds is probably Middle Devonian in age and related to the Acadian Orogeny. Localized intense faulting, thrusts and isoclinal folds occur adjacent to the Chedabucto Fault and may be related to the docking of the Meguma Terrane to the south.

PLUTONIC ROCKS

Introduction

The plutonic rocks of the Antigonish Highlands range from late Precambrian to Devono-Carboniferous in age (Fig. 10). The late Precambrian plutons include the gabbroic Greendale Complex and nearby alaskite in the northern Antigonish Highlands, and the gabbroic to dioritic Eden Lake and Black Brook complexes in the southern highlands (Fig. 3). Grani-

toid rocks of late Precambrian or Cambrian age include the James River and Ohio plutons which occur in the eastern highlands (Fig. 3). Unnamed gabbroic stocks in the northemmost Antigonish Highlands are interpreted to be Early Cambrian in age (Fig. 3). Their close spatial association and similar chemistry to the Cambrian Arbuckle Brook Formation mafic volcanic rocks suggest a genetic association. Cambrian or later bodies include a number of small syenitic and granitic stocks which intrude the Cambro-Ordovician rocks. Devono-Carboniferous granitoid plutons include the Gunn Lake Pluton in the southern highlands, the Barneys River plutons in the central highlands and a number of small unnamed plutons.

Late Precambrian plutons

Greendale Complex

Field relationships. The Greendale Complex has a strong magnetic signature (Geological Survey of Canada, 1982) which defines a roughly circular shape approximately 3 km in diameter, truncated by the present coastline (Fig. 4; Map 1749A). This signature suggests that the overall shape of the intrusion may have been cylindrical, although this may be an oversimplification since dykes up to 20 m in width occur within the complex. The complex is posttectonic with respect to the penetrative structures observed in the host rocks (the Georgeville Group), because not only do contacts cut across these structures but contact metamorphic porphyroblasts grow over the structural fabrics.

The intrusive representatives of the complex show remarkable textural, mineralogical and lithological similarities to the appinite suite as described by Pitcher and Berger (1972) and Bowes and MacArthur (1976). The most important field characteristics of these rocks, which are typical of appinite suites, are:

1) the overwhelming dominance of amphibole relative to other mafic phases;
2) the rapid and irregular textural variations from coarse pegmatitic to relatively fine grained poikilitic or porphyritic amphibole; this may occur on the scale of an outcrop or a hand specimen;
3) although the intrusion is predominantly mafic in character, there is abundant evidence of late stage felsic magmas associated with the complex (Fig. 63 and 64).

The complex may be subdivided on the basis of texture and mineral assemblage into eight distinct lithologies (the approximate surface area percentage abundances are in parentheses): marble (10) and volcanic host rock (10), coarse hornblende pegmatite (12), plagioclase porphyritic rock (30), hornblende plagioclase porphyry (8), hornblende porphyritic rock (8), hornblende poikilitic rock (10), lamprophyre (5) and felsic rock (7). Although much of the igneous rocks resemble hornblende diorites, geochemically they are gabbroic with some intermediate tendencies (below).

All lithologies except for the lamprophyres and hornblende-poikilitic gabbros occur in the northeastern part of the complex (Fig. 65a). Contacts between the units may be gradational or sharp and irregular or planar. Intrusion/wall-rock contacts occur irregularly in the complex. Host rocks may be either rotated blocks or in-situ roof pendants. Marble is buff coloured, saccharoidal and displays banding parallel to bedding. Basalt may be present as relict cores up to 4 m in diameter surrounded by intrusive rock (usually pegmatite), or as tabular bodies interbedded with the marble but with gradational contacts with the intrusive rocks (usually with fine grained porphyries). Pegmatites (with hornblende grains up to 10 cm in length) occur not only as "shells" enclosing basaltic host rocks but also as discontinuous lenses and veins in the contact zones, and as discrete tabular bodies. Contacts between the pegmatites and the fine grained porphyries are commonly gradational. Fine grained lithologies are dominant in the core of the pluton, where they occur as multiple sheets up to 20 m in width. Rounded xenoliths are common. The fine grained porphyry generally has gradational contacts

Figure 63. Felsic "sweaty" pods within a fine grained appinite. GSC 204717.

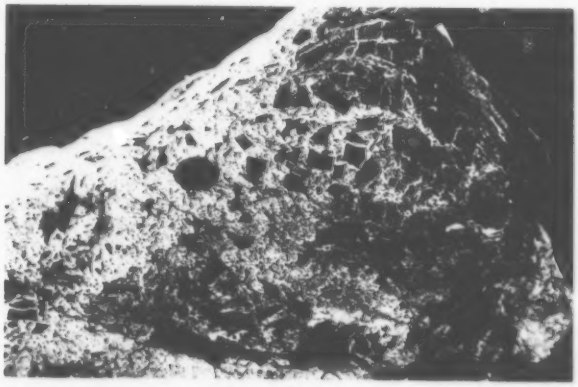


Figure 64. Boulder showing evidence of in situ brecciation. This happens where the fluid pressure locally exceeds the lithostatic pressure and can overcome the shear strength of the rock. This feature is typical of appinite suites (e.g. Bowes and MacArthur, 1976). GSC 204717A.

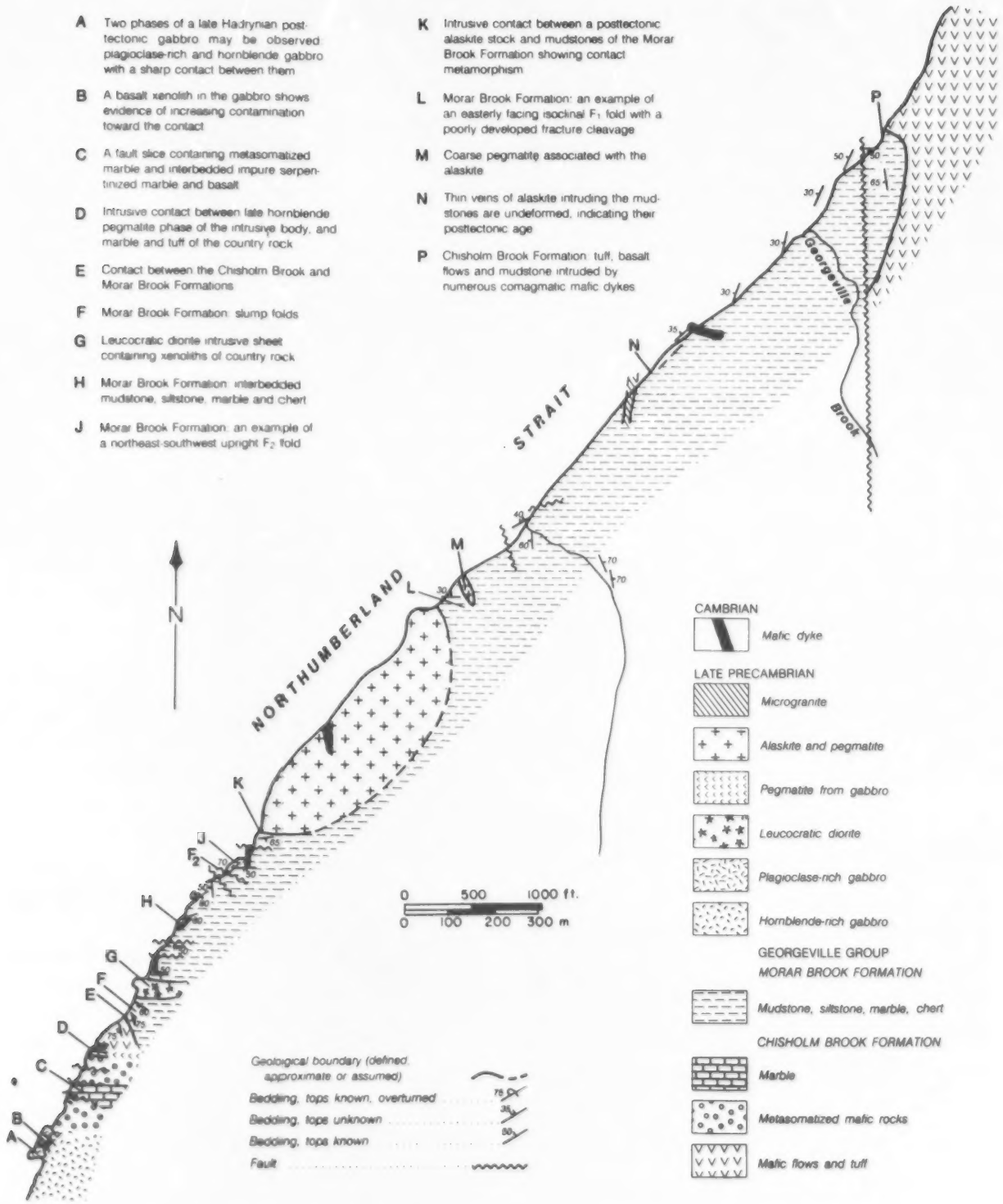

A Two phases of a late Hadrynian post-tectonic gabbro may be observed: plagioclase-rich and hornblende gabbro with a sharp contact between them
B A basalt xenolith in the gabbro shows evidence of increasing contamination toward the contact
C A fault slice containing metasomatized marble and interbedded impure serpentinized marble and basalt
D Intrusive contact between late hornblende pegmatite phase of the intrusive body, and marble and tuff of the country rock
E Contact between the Chisholm Brook and Morar Brook Formations
F Morar Brook Formation: slump folds
G Leucocratic diorite intrusive sheet containing xenoliths of country rock
H Morar Brook Formation: interbedded mudstone, siltstone, marble and chert
J Morar Brook Formation: an example of a northeast-southwest upright F_2 fold
K Intrusive contact between a posttectonic alaskite stock and mudstones of the Morar Brook Formation showing contact metamorphism
L Morar Brook Formation: an example of an easterly facing isoclinal F_1 fold with a poorly developed fracture cleavage
M Coarse pegmatite associated with the alaskite
N Thin veins of alaskite intruding the mudstones are undeformed, indicating their posttectonic age
P Chisholm Brook Formation: tuff, basalt flows and mudstone intruded by numerous comagmatic mafic dykes
N
NORTHUMBERLAND STRAIT
Georgeville Brook
0 500 1000 ft.
0 100 200 300 m
Geological boundary (defined, approximate or assumed)
Bedding, tops known, overturned
Bedding, tops unknown
Bedding, tops known
Fault
CAMBRIAN
Mafic dyke
LATE PRECAMBRIAN
Microgranite
Alaskite and pegmatite
Pegmatite from gabbro
Leucocratic diorite
Plagioclase-rich gabbro
Hornblende-rich gabbro
GEORGEVILLE GROUP
MORAR BROOK FORMATION
Mudstone, siltstone, marble, chert
CHISHOLM BROOK FORMATION
Marble
Metasomatized mafic rocks
Mafic flows and tuff

Figure 65a. Detailed geological map and descriptions of the shoreline section of the Greendale Complex. Locations shown in Figures 4 and 65b.

with the pegmatites and host rocks, but sharp contacts with other fine grained porphyries, lamprophyres and felsic veins. It consists almost totally of hornblende and plagioclase (Fig. 66b). Hornblende varies in habit from interstitial crystals to euhedral prisms up to 30 mm in length. Plagioclase forms discrete crystals up to 10 mm in length, commonly has a highly irregular habit and may enclose cores of patchy material giving it a "dirty" appearance. Plagioclase also occurs in irregular segregations within the porphyry.

Felsic rocks (lenses, microgranitic and aplitic veins and dykes) constitute the latest igneous rocks in the Greendale Complex. Lenses up to 1 m in length are common in all other lithologies (Fig. 63). Contacts with other lithologies may be sharp or gradational. Felsic veins and dykes up to 3 m in width generally have sharp contacts and generally occur adjacent to intrusive rock contacts. They are locally inhomogeneous; for example, a vein may vary laterally from a biotite-bearing microgranite to a quartz-feldspar pegmatite.

Lamprophyres and hornblende-poikilitic gabbro do not occur within the area mapped in Figure 65a, but are common in the southwestern parts of the complex (Fig. 65b). Lamprophyres are black and occur as dykes 2 to 4 m wide with irregular but sharp contacts with all other lithologies, or as ellipsoidal pods up to 0.8 m in length (Fig. 66a).

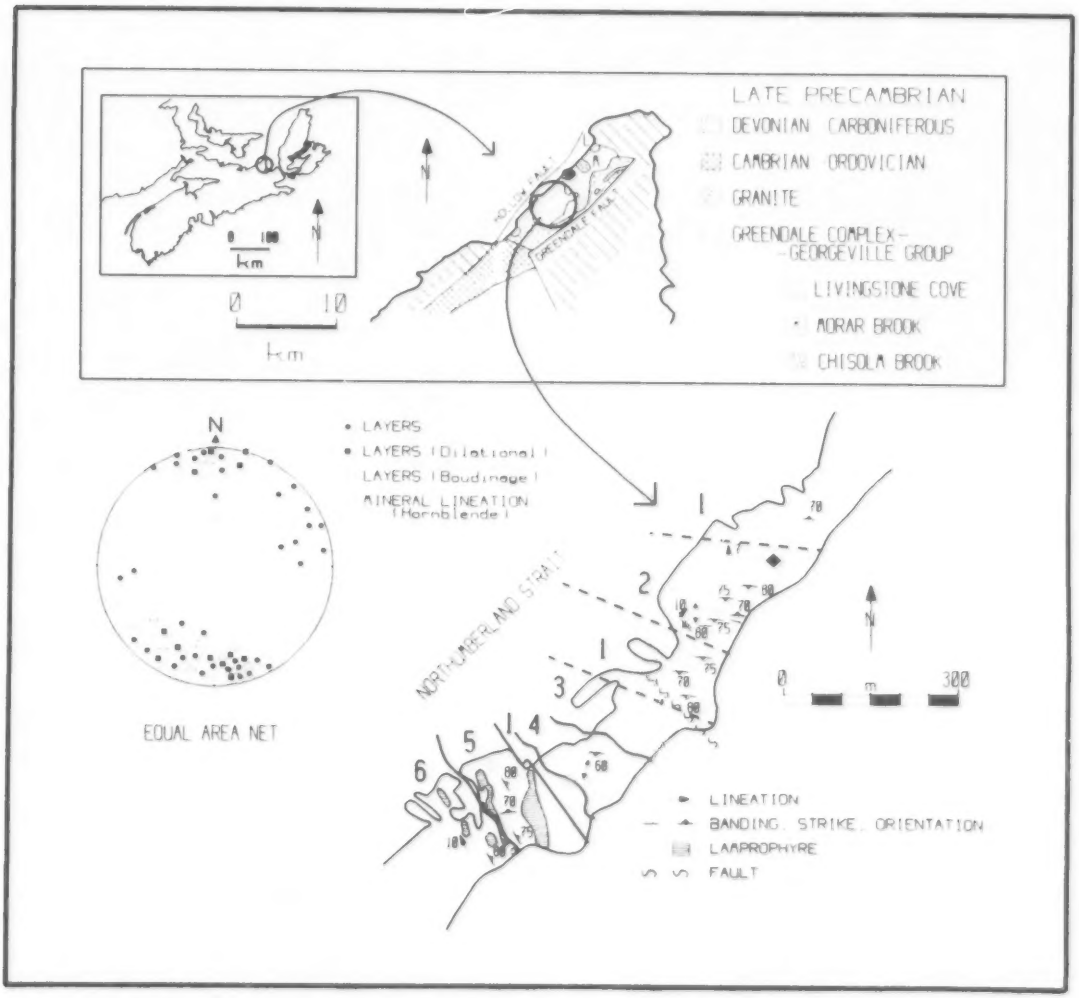


Figure 65b. Summary map of the layered portion of the Grendale Complex. Inset shows location and local geology. Stereoplot is an equal area projection showing structural data. 1=fine grained porphyry, 25% felsic veins; 2=interlayered hornblende and plagioclase porphyries, 15% felsic veins; 3=host rock hornfels; 4=poikilitic hornblende porphyry, 15-20% felsic veins; 5=interlayered porphyry and lamprophyre, 20% felsic veins; 6=porphyry with lamprophyric boundins.

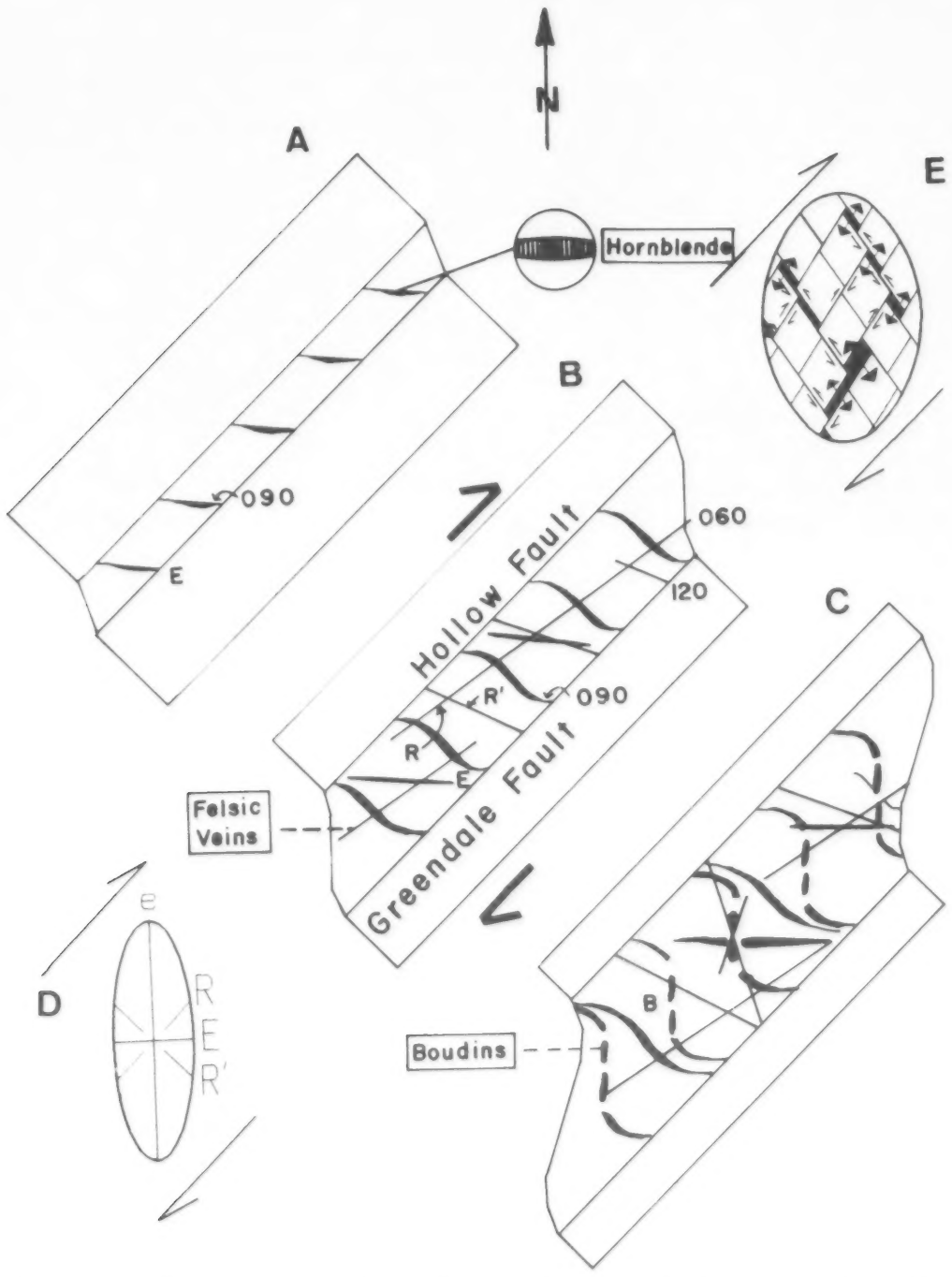


Figure 65c. Schematic diagram showing development of layering in Greendale Complex. Thick lines are layers; thin lines are felsic veins. Scheme modified from Ramsay and Huber (1983). (a) Layers initially develop parallel to extensional plane in instantaneous strain ellipse at approximately 90°, hornblende crystals elongate perpendicular to margins of layers. (b) Conjugate felsic veins occur with large dihedral angle and intrude parallel to shears (R-R'). Early layers are rotated toward plane of flattening. Instantaneous strain ellipsoid shows expected orientations of planes of extension (E), synthetic shears (R), antithetic shears (R'), plane of flattening, and extension lineation (e in D), assuming dextral movement on Hollow and Greendale faults. (c) Progressive dextral shear rotates early layers and felsic veins toward plane of flattening (approximately 160°) and results in new 090° layers. (d) Schematic representation of how conjugate vein system could develop by bulk inhomogeneous flattening due to combination of displacement parallel to and opening at high angle to length of individual shears (after Ramsay and Huber, 1987).

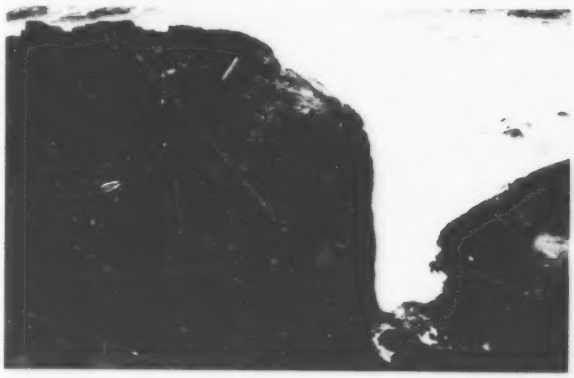


Figure 66a. Lamprophyric clots up to 1 m in length (central part of the photograph) occur within a fine grained appinitic host. A more hornblende-rich appinite intrudes this sequence (left). GSC 204717C.

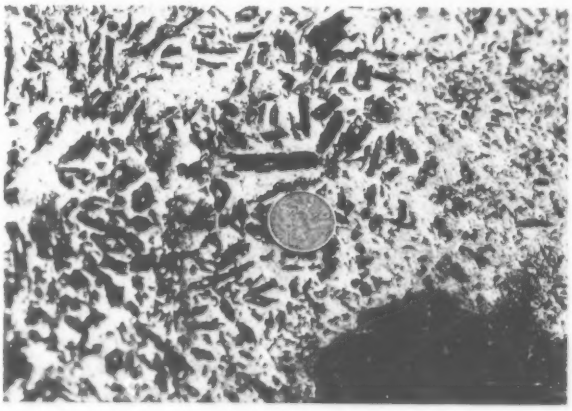


Figure 66b. Typical fine grained hornblende porphyry. GSC 1991-187.

Figure 66c. Example of 090° layering. Note orientation of hornblende crystals perpendicular to the walls. GSC 1991-184.

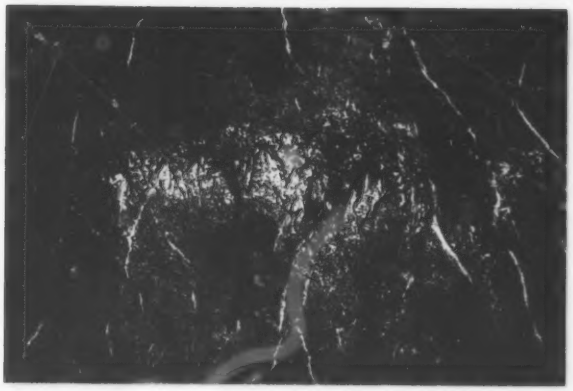


Figure 66d. Pegmatite, with hornblende growing perpendicular to the walls. GSC 1991183.

Figure 66e. Example of boudinage in 160° layers. GSC 1991-182.

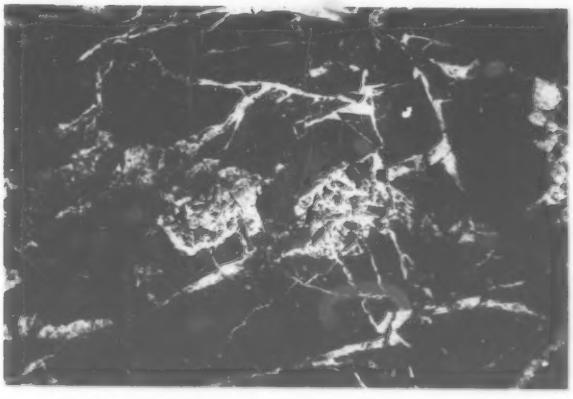


Figure 66f. Felsic veins in conjugate sets. GSC 1991-186.

Hornblende-poikilitic gabbro consists of thick sheets containing hornblende (up to 10 cm in length) poikilitically enclosing plagioclase and augite. Contacts with the fine grained porphyries are gradational. There is abundant evidence that the complex is exposed near its roof. There are abundant inclusions of host rock, some of which may be roof pendants. Further, both mafic and felsic pegmatites are common. The presence of hornblende and absence of olivine and pyroxene in gabbro melts have been attributed to high $p(H_2O)$ conditions (Yoder and Tilley, 1962) that commonly occur near the roof of gabbros (Bowes and MacArthur, 1976).

Layering. All lithologies display spectacular interlayering, particularly on the shoreline section. The layers are generally defined by variations in grain size and/or relative abundance of hornblende and plagioclase. In some of the layers, hornblende is aligned orthogonal to the contact (Fig. 66c). Gabbroic pegmatites occur as dykes up to 10 m wide or as thin layers up to 1 m wide that are generally parallel to layering and contain hornblendes up to 30 cm long that are commonly aligned orthogonal to the contact with the adjacent rocks (Fig. 66d). Lamprophyres, containing about 90% hornblende poikilitically enclosing olivine, augite, and magnetite, occur in 2 to 6-m-wide dykes. They also occur adjacent to these dykes as discontinuous layers and ellipsoidal pods within the hornblende-plagioclase porphyries. The pods are probably boudins (Fig. 66e). Fine grained felsic rocks occur typically in conjugate veins (Fig. 66f) that intrude all the lithologies, indicating that they were emplaced late in the history of the complex. These veins commonly terminate in felsic pods that display pegmatitic textures and have gradational contacts with the more mafic rock types. These relations indicate that the emplacement of felsic rocks was at a late stage and cogenetic with the development of the Greendale Complex. This is supported by the similar range in initial $^{87}Sr/^{86}Sr$ ratios (0.7020 to 0.7045) in both mafic and felsic rocks in the complex (R. F. Cormier, pers. comm., 1989).

Layering occurs predominantly in the central part of the complex where all the above rocks (with the exception of the felsic rocks) are interlayered (Fig. 65b). These layered zones, which may be up to 150 m wide, consist entirely of steeply dipping layers 5 mm to 50 cm wide. Contacts between layers are sharp, and although on an outcrop scale they are broadly planar, they may be locally irregular and/or crosscutting. The felsic rocks occur in conjugate sets at variable angles (ranging from 5° to 40° to the layering). Contact relations between more mafic lithologies are complex on all scales, and no consistent age relations between rock types can be determined. The contacts are generally sharp, but locally irregular. The field relations indicate that the complex was probably formed by multiple intrusive phases of predominantly mafic, but chemically varied magma from a subjacent magma chamber.

Orientation of Layering. Layering within the complex strikes from south-southeast to east and is generally vertical or steeply dipping (Fig. 65b). In places, hornblende crystals are aligned perpendicular to the layers (Fig. 66c). The layering is spectacularly enhanced by thin layers of coarse hornblende pegmatite; hornblende crystals as long as 50 cm define a strong lineation generally perpendicular to the margins of the pegmatite. The layering defined by the pegmatites is always parallel or subparallel to the finer grained layers in the same outcrop. The strike of these layers is east-northeast to east-southeast with a cluster at around 090° (Fig. 65b). Lineations defined by hornblende in these layers are subhorizontal and generally trend south-southeast.

Although the layering in the complex strikes dominantly between 060° and 100°, there are regions in which it is as much as 140° or 160°. In these regions many of the layers display boudinage (Fig. 66e). Discontinuous layers of lamprophyre that occur adjacent and parallel to lamprophyric dykes are also interpreted as boudins of lamprophyric dykes. Layers that display boudinage typically strike south-southeast to south (Fig. 65b), with a cluster at around 160°. In thin section, these layers show subhorizontal alignment of large to small hornblende and plagioclase crystals parallel to their boundaries. Although the coarse crystals are angular, the fine crystals are thin and highly elongate, have rounded lobate terminations, and may have been plastically deformed on a grain scale. Pegmatites occur parallel to these layers, although they are less abundant than those associated with the 090° layers and the hornblendes within them show no consistent orientation.

Felsic veins occur in conjugate sets. The dihedral angles between vein sets vary. Conjugate felsic veins that intrude the 090° layers have large dihedral angles (between 50° and 80°) are steeply dipping, and have strikes clustering at about 060°. Veins have evidence of dextral offset, whereas some 120° veins display sinistral offset. Felsic veins that intrude the 160° layers occur in two conjugate sets; an early set with a small dihedral angle (i.e. <30°) that is subparallel to the 160° layers, and a later set with a large dihedral angle that intrudes the early set (see Fig. 65c) and is similar in orientation to the conjugate set that intrudes the 090° layers.

Origin of Layering. The character and orientation of layering in the Greendale Complex are consistent with the stress field to be expected in a region undergoing transpression due to dextral shear on a set of northeast-trending faults (Fig. 65c; modified after Ramsay and Huber, 1983). It has been shown (e.g., Ramsay and Graham, 1970) that slip along a set of faults generates a simple shear system between these faults. This system should give rise to north-trending extension. There is abundant evidence for subhorizontal extension in a broadly north-south direction during emplacement and solidification of the complex. This extension is inferred to result in the steeply dipping east-striking layers with subhorizontal hornblende crystals trending north. The hornblende growth is attributed to development from a substrate into a zone of low pressure. These relations indicated a dilational origin for the layering in the pegmatites and, given the similar orientation, probably for layering in the finer grained rocks. Thus, the 090° layering is interpreted to be the result of progressive and multiple injection of appinitic magma from either an interstitial liquid or a subjacent body of magma into fractures developing parallel to the extensional plane of the instantaneous

strain ellipsoid. Extensional fractures may have developed within the crystallizing layers when they became capable of transmitting stress (i.e., when they became approximately 70% crystalline; Hibbard and Watters, 1985). The varied texture (pegmatitic versus fine grained) may reflect variations in H_2O pressure in the invading magma as described by Jahns and Burnham (1969).

The development of boudinage in approximately 160° layers is also consistent with north-south extension. Progressive regional dextral shear should give rise to clockwise rotation of early-formed east-striking layers (Fig. 65c). Both the presence of layers with southerly strikes and their tendency to exhibit boudinage may be explained by this process because rotation moves them toward the plane of flattening for the instantaneous strain ellipsoid (Fig. 65c). The south-striking layers are therefore interpreted as early-formed layers that underwent rotation and subsequent layer-parallel extension during progressive dextral shear. The mineral-grain fabrics at the margins of some of the 160° layers are consistent with this interpretation; they may reflect plastic deformation on the grain scale during progressive shear associated with the rotation.

The orientations of the felsic veins are also consistent with the proposed stress field. The predominant conjugate set with strikes of 060° and 120° conforms in both its attitudes and offset characteristics (dextral for 060°, sinistral for 120°) to the planes expected for shear failure in Coulomb-Navier theory (see D, Fig. 65c). The smaller dihedral angle associated with the 160° layers may be explained by rotation toward the plane of flattening during progressive dextral shear. The apparent anomaly of dilation across proposed shear-failure planes may reflect alternation of displacement between shear planes of a given conjugate set (Ramsay and Huber, 1983) and along which there are small changes in attitude, as proposed by Reches (1978).

Summary. Structures in the Greendale Complex are compatible with control by a stress field imposed by dextral motions on northeast-trending faults that bound the Antigonish Highlands. Direct evidence for this stress field is provided by large-scale structures developed in the Georgeville Group shortly before emplacement of the Greendale Complex. The Greendale Complex is thought to have been emplaced by progressive and repeated extensional failure of the roof of the magma chamber, which must have initially consisted of Georgeville Group country rocks but was gradually replaced by frozen parts of the intrusive complex. The attitudes of the planes of extensional failure were governed by the (constant) attitude of the stress field, and imparted a strong, subvertical east-trending layering to the complex. In places, however, the complex appears to have undergone progressive plastic strain in response to the stress field, giving rise to clockwise rotation of the layering. The felsic veins appear to represent a different mode of failure (shear in contrast to extensional failure) at a late stage in the evolution of the complex, which may be due to cooling of the roof of the magma chamber. The markedly different character of these late rocks, despite their apparent comagmatic evolution, indicates profound fractionation in the waning stages of evolution of the magma.

This study demonstrates the importance of considering regional and local tectonic setting in the study of igneous fabrics such as layering. There is little in the observed internal structure of the Greendale Complex that is not attributable to the external stress field. This may, however, reflect the chance intersection of the erosion surface with the roof of the magma chamber rather than the characteristics of the whole complex.

Petrography. Rocks of the intrusive suite may be termed "fresh" or "altered" on the basis of the degree of preservation of hornblende. "Fresh" fine grained (<15 mm) porphyritic rocks consist of variable modal proportions of fresh hornblende (some with actinolitic rims) and plagioclase with accessory quartz, sphene and magnetite. In these rocks, plagioclase is invariably altered to epidote or sericite or both. In "altered" porphyries, hornblende is altered to chlorite or tremolite. Chlorite may contain inclusions of zircon. Plagioclase may have fresh albitic overgrowths enclosing a core of epidote and/or sericite, or, more rarely, calcite. The altered cores of the plagioclases are pseudomorphs after euhedral feldspar, whereas the overgrowths are irregular in form.

Hornblende pegmatites consist of coarse, euhedral to ragged, prismatic hornblende up to 15 cm in length (with locally developed hornblende-actinolite zoned couplets) enclosing patches of felsic material and sericitized plagioclase. Incipient secondary alteration of hornblende to tremolite or chlorite occurs locally. Plagioclase is up to 6 mm in length

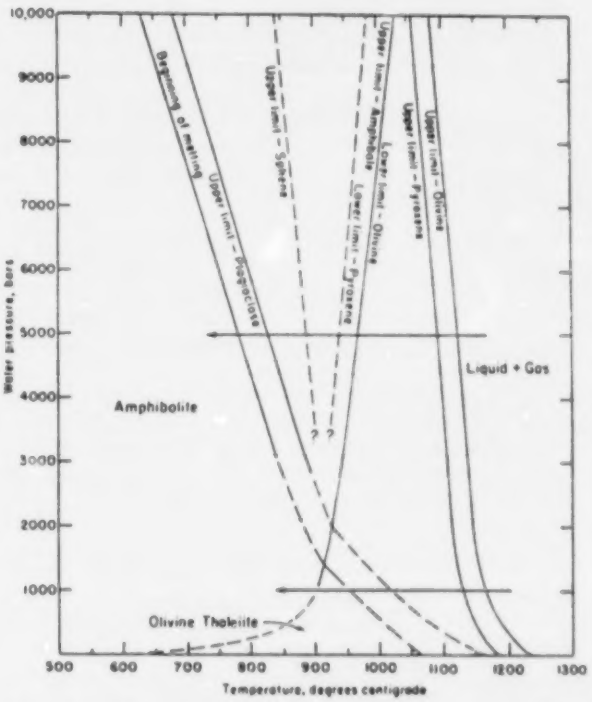


Figure 67. Phase relationships (after Yoder and Tilley, 1962) for olivine tholeiite as a function of $p(H_2O)$.

and is replaced as in the altered porphyries. The matrix of some of the pegmatites may be almost exclusively composed of calcite. Other pegmatites have a matrix consisting of variable proportions of plagioclase, epidote, sphene, apatite and chlorite.

Hornblende-poikilitic gabbro is characterized by xenomorphic hornblende containing inclusions of plagioclase, augite, serpentine pseudomorphs of olivine and sphene. Lamprophyres consist of hornblende up to 6 cm in length enclosing cumulus grains of olivine, augite and minor hypersthene. Biotite, epidote, calcite, plagioclase and rutile may be present as minor accessory phases.

In most lithologies, hornblende commonly encloses plagioclase suggesting that it commenced crystallizing after plagioclase. This relationship allows an estimate of the pressure of the intrusive event (Fig. 67; Yoder and Tilley, 1962). Assuming $p(H_2O) = p(total)$, then the maximum pressure is about 2 kb.

Microgranites are allotriomorphic equigranular rocks consisting of quartz, potassium feldspar, plagioclase, muscovite, chlorite after hornblende, minor zircon and sphene. Aplites have a similar mineralogy and texture, but are finer grained. Aplites may also contain minor garnet. Felsic "sweaty" pods consist of quartz and potassium feldspar ± plagioclase.

Contact metamorphism. The marbles and impure marbles adjacent to the Greendale Complex contain the following assemblages: (i) calcite-wollastonite-serpentine-quartz ± pyrite; (ii) calcite-diopside-quartz ± pyrite; and (iii) calcite-tremolite-epidote-quartz ± pyrite. Fine equigranular calcite generally accounts for between 65 and 75 volume percent of the rock. Tremolite occurs in minute and randomly oriented felted crystals. Talc forms ragged platy crystals that in part may pseudomorph diopside, or more rarely, hornblende. Serpentine pseudomorphs forsterite, although forsterite may be preserved in the cores of some of these crystals.

Basalts in the contact aureole display a wide variety of textures and mineralogy. In some cases, variable textures may occur on the scale of a thin section. Albite, quartz and epidote are ubiquitous. The remainder of the paragenesis may consist of: (i) calcite-chlorite-sphene ± ilmenite; (ii) calcite-chlorite-tremolite-sphene ± magnetite; (iii) calcite-chlorite-hornblende-sphene ± magnetite; (iv) tremolite-hornblende-sphene ± magnetite; or (5) chlorite-tremolite ± magnetite ± sphene. Forsterite and diopside, preserved in one thin section, represent the highest grade of metamorphism associated with the intrusion.

Albite occurs as stubby porphyroblasts (less than 3 mm in length), is commonly sericitized and may contain inclusions of calcite, epidote or chlorite. Quartz is anhedral and interstitial. Epidote occurs as small anhedral globular blebs. Hornblende may be partly to completely pseudomorphed by chlorite and/or tremolite. It commonly contains inclusions of sphene. Calcite is irregularly distributed, fine grained and extremely ragged in morphology. Rutile, where present, is usually in trace amounts or associated with pyrite and magnetite in veins. Magnetite may also occur in thin stringers. Apatite and prehnite are minor accessory minerals; prehnite is generally found only in veins. The highest temperature assemblages immediately adjacent to the Greendale Complex and in xenoliths are indicative of a pyroxene hornfels facies of metamorphism. This decreases to the hornblende hornfels facies a short distance from the contact.

Table 8. Geochemistry of the fresh fine grained appinitic rocks of the Greendale Complex

	E16-683	E16-689	E16-791	E16-003	E16-048	E16-080	E16-192A	E16-192B
SiO_2	51.1	48.0	49.7	50.6	49.0	49.8	51.2	53.0
TiO_2	1.7	1.5	1.6	1.4	1.8	1.6	1.5	1.4
Al_2O_3	17.5	15.9	15.5	20.7	16.8	19.4	17.2	19.8
FeO*	11.2	9.7	11.7	7.8	10.2	10.5	10.5	8.9
MgO	6.5	10.9	8.2	5.7	8.4	5.2	5.5	4.5
CaO	8.2	9.1	9.6	8.2	9.8	8.4	8.6	7.5
Na_2O	3.3	2.9	3.1	3.3	3.0	3.8	3.3	4.1
K_2O	0.7	1.7	0.5	2.2	1.1	1.1	1.0	1.2
P_2O_5	0.2	0.4	0.2	0.2	0.2	0.4	0.3	0.4
Rb	19	50	13	105	31	34	20	25
Sr	294	297	222	434	212	501	418	464
Y	31	33	25	27	23	24	30	28
Zr	72	43	43	46	88	53	59	58
Nb	<5	<5	<5	<5	<5	<5	<5	<5
Ni	57	78	71	7	n.d.	n.d.	8	32
Cr	90	62	63	23	n.d.	n.d.	133	128

n.d. = not determined

Age. Field relationships constrain the age of the Greendale Complex. This constraint is provided by its crosscutting relationships not only with the host rocks of the Georgeville Group but also with the penetrative fabrics in these rocks. A $^{40}Ar/^{39}Ar$ isotopic analysis of hornblende from a hornblende pegmatite yielded a more precise age determination. Two samples yielded plateaus with ages of 611.2 ± 3.8 Ma and 620.3 ± 4.8 Ma (Keppie et al., 1990).

Geochemistry. The interpretation of the geochemistry of the Greendale Complex is complicated by the extensive alteration observed in thin section and in the chemistry. The petrology has been examined in detail by Murphy (1982) and is summarized here. Geochemical analyses of the Greendale Complex are presented in Table 8. For example, TiO_2 varies from 1.4 to 1.7 weight percent and Zr varies from 43 to 88 ppm. Ti displays a positive correlation with Fe, Mg, Cr and Ni and a negative correlation with Si, Na, K, Rb and Zr. Most of the variation in elemental content is inferred to be the result of element mobility at temperatures between the liquidus and the solidus and involved mass transfer in a magma rich in H_2O and CO_2. CO_2 was probably derived from an assimilation of marble xenoliths from the Chisholm Brook Formation. The presence of these volatiles in the magma led to the partitioning of Si, K, Na, Rb and Zr into the H_2O-rich silicate melt and enrichment in the centre of the body (Murphy, 1982). On the other hand, Ti, Fe, Mg, Ca, Sr, Y, Ni and Cr partitioned into the CO_2-rich melt and fluid phases and were driven toward the margins of the body where they were deposited at the edge of the transport system in the marginal pegmatites and adjacent basalts. A later subsolidus metasomatism localized along the pluton margin generally accentuated or did not affect earlier alteration trends, although metasomatism occasionally reversed earlier trends (Murphy, 19[illegible]).

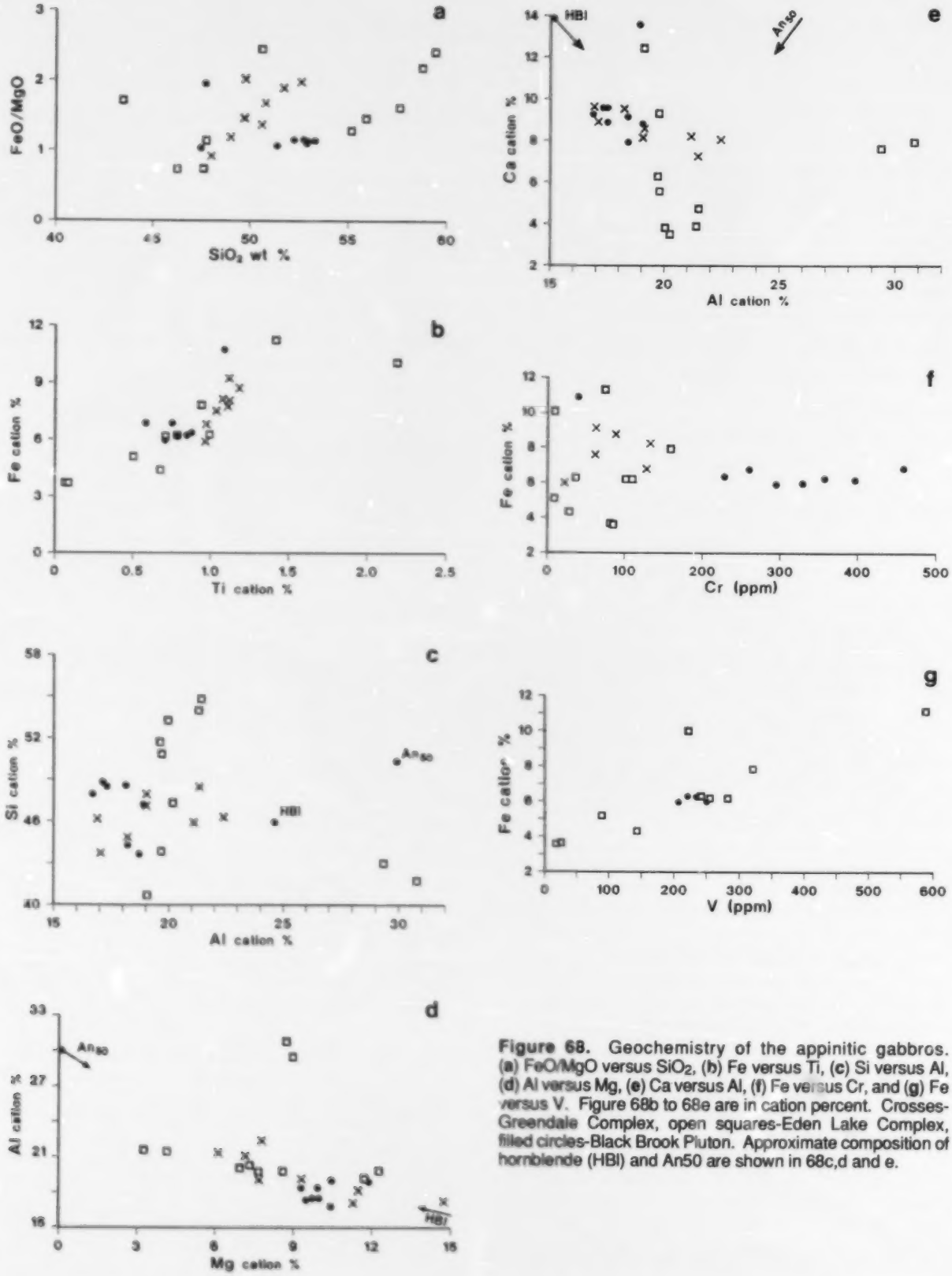


Figure 68. Geochemistry of the appinitic gabbros. (**a**) FeO/MgO versus SiO_2, (**b**) Fe versus Ti, (**c**) Si versus Al, (**d**) Al versus Mg, (**e**) Ca versus Al, (**f**) Fe versus Cr, and (**g**) Fe versus V. Figure 68b to 68e are in cation percent. Crosses-Greendale Complex, open squares-Eden Lake Complex, filled circles-Black Brook Pluton. Approximate composition of hornblende (HBl) and An50 are shown in 68c,d and e.

The extent of the alteration makes it extremely difficult to determine the original magmatic character of the Greendale Complex. If one considers only the "fresh" rocks it is clear that the complex is bimodal with a silica gap between 46 percent and 58 percent, a feature typical of appinite suites (Hall, 1967; Pitcher and Berger, 1972). The ubiquitous presence of magmatic oxyhornblende is attributed to the high partial pressure of H_2O which causes hornblende to be crystallized at the expense of olivine and pyroxene (Fig. 67; Yoder and Tilley, 1962). However, the presence of olivine and pyroxene poikilitically enclosed in hornblende in lamprophyre dykes and inclusions and the hornblende gabbro indicate less hydrous conditions in some parts of the intrusive body. These factors suggest that the parental magma was basaltic. The presence of olivine and hypersthene suggest it may have been tholeiitic. The increase of FeO/MgO with SiO_2 and the positive correlation between Fe and Ti is consistent with this relationship (Fig. 68a,b). Variations in Si and Al (Fig. 68c) can be attributed to variations in the modal abundances of hornblende and plagioclase. Negative Al-Mg and especially Ca-Al relationships (Fig. 68d,e) are consistent with hornblende fractionation.

Eden Lake Complex

Field relationships. The Eden Lake Complex occurs in the southernmost Antigonish Highlands adjacent to the Chedabucto Fault (Fig. 3). The complex consists of four small bodies of between 3 and 6 km in surface dimension. These bodies are separated by north-northwest-trending faults, by Ordovician-Silurian rocks or by Pleistocene gravel. The roughly circular shape suggests that the form of the intrusion may have been a cylindrical stock.

The Eden Lake Complex intrudes Precambrian rocks and is unconformably overlain by an Upper Ordovician-Silurian sequence. Although the contact is not exposed, the complex apparently truncates the contact between the Frasers Brook and Moose River members of the late Precambrian Keppoch Formation. On the other hand, arkoses of the Bears Brook Formation unconformably overlie this complex. Although there is a gap in exposure of about 5 m between the pluton and the arkoses, there is no sign of contact metamorphism. Bedding in the arkoses is approximately parallel to the contact.

The western localities are separated from the eastern localities by the north-northwest-trending Moose River Fault, the north-trending Garden River Fault and by spatially-associated Quaternary sand and gravel deposits.

Age. On the basis of field relationships, the Eden Lake Complex could be late Precambrian to Late Ordovician in age. However, a K-Ar determination on hornblende yielded an age of 582 ± 32 Ma (Wanless et al., 1967), suggesting a latest late Precambrian or Early Cambrian age (see Palmer, 1983). The Eden Lake Complex is remarkably similar to the Greendale Complex and phases of the Black Brook pluton, which also posttectonically intrudes the Georgeville Group and yield ca. 620-600 Ma ages. The Eden Lake and Greendale complexes and Black Brook pluton are inferred to be broadly coeval and are late Precambrian in age. The discrepancy in ages may be due to the different dating techniques used.

Field description. The Eden Lake Complex is dominated by medium to fine grained equigranular hornblende-plagioclase-bearing gabbros and diorites. These lithologies make up approximately 90 percent of the surface exposure and thus are similar to the dominant lithologies of the Greendale Complex. Coarse hornblende-bearing pegmatitic rocks occur in pods. In comparison with the Greendale Complex, the pegmatites are relatively minor in abundance. The contacts between the pegmatites and the medium grained host rock are irregular but sharp. Felsic rocks are heterogeneously distributed throughout the complex as small (i.e. less than 1 m in diameter) "sweaty" pods or as irregular veinlets. The following characteristics of the Eden Lake Complex are typical of appinite suites (see Bowes and MacArthur, 1976): (i) the overwhelming dominance of hornblende as a mafic phase; (ii) the lack of pyroxene and olivine in rocks of broadly

gabbroic chemistry; (iii) the close spatial association of fine grained hornblende-plagioclase rocks with hornblende-plagioclase pegmatites; (iv) the occurrence of late stage felsic "sweat pods" and veinlets; and (v) the heterogeneous distribution of the lithologies on a regional, outcrop and hand specimen scale.

Petrography. The fine grained rocks are dominated by hornblende (minor actinolite) and plagioclase and their alteration products, with minor quartz, opaques and apatite. Hornblende occurs as euhedral to subhedral grains up to 3 mm in length. It is generally green to green-brown in colour. Rarely, the amphibole displays oscillatory zoning with alternating zones of actinolite and hornblende composition. Hornblende may show incipient (?) to complete alteration to chlorite and actinolite, and more rarely, biotite.

Plagioclase occurs as stubby grains up to 4 mm in length. It is generally albitic in composition and is commonly extensively saussuritized to epidote and mica. Quartz varies considerably in abundance occurring up to 10 modal percent in the more dioritic rocks and is typically interstitial. Opaques consist of equant grains (up to 1 mm) of magnetite (partly altered to hematite) and minor pyrite. Apatite occurs as slender prisms up to 0.5 mm in length.

Geochemistry. The Eden Lake Complex has a relatively heterogeneous chemistry (Table 9). SiO_2 abundances indicate that most of the complex is mafic to intermediate in composition (Fig. 68a) ranging (on a volatile-free basis) from 43 to 60 weight percent. The complex contains about 10 percent by volume of felsic rocks; their chemistry is reported in Table 9b and is not discussed further here. With the exception of two samples that have relatively coarse grained plagioclase, Al-Mg and Ca-Al relationships (Fig. 68d,e) are consistent with hornblende fractionation. The positive

Table 9a. Geochemistry of the mafic volcanic rocks of the Eden Lake Complex

	THO3-1F	7A	7B1	7C	7E1	7F	8C1	8C2	8E	8F1
SiO_2	56.95	53.78	58.39	45.61	41.13	53.81	44.71	46.24	47.94	55.11
TiO_2	0.68	0.99	0.95	1.29	1.85	1.13	0.13	0.11	2.91	1.34
Al_2O_3	19.08	17.69	19.37	17.40	16.35	17.32	27.91	26.84	17.38	17.54
Fe_2O_3	7.17	8.78	6.23	10.96	15.19	8.60	5.17	5.32	13.62	8.71
MgO	2.93	6.10	2.31	8.61	7.97	5.32	6.25	6.51	4.99	4.80
MnO	0.17	0.17	0.19	0.27	0.27	0.18	0.32	0.31	0.39	0.16
CaO	3.85	5.57	4.78	9.12	11.77	6.10	7.98	7.65	3.45	3.71
Na_2O	4.26	2.91	4.07	2.46	1.15	2.83	1.56	2.13	3.63	3.63
K_2O	2.07	2.14	2.33	0.84	0.60	1.59	3.18	2.54	1.09	1.13
P_2O_5	0.40	0.20	0.20	0.24	0.10	0.23	0.09	0.08	0.72	0.34
LOI	2.46	2.54	1.23	2.30	2.56	2.69	4.31	4.14	3.46	3.92
TOTAL	100.01	100.87	100.04	99.11	98.95	99.90	101.62	101.98	99.57	100.38
Rb	53	56	63	19	16	41	96	79	26	33
Sr	617	297	468	535	568	370	796	859	287	351
Y	25	21	18	17	12	24	6	6	42	30
Zr	157	93	428	62	37	101	23	23	262	252
Na	7	6	6	4	5	7	4	3	31	9
Ba	935	856	804	376	229	503	1263	870	574	580
Ni	tr	39	5	63	75	20	73	71	28	32
Cr	9	110	29	160	74	100	85	83	10	37
V	89	253	141	321	594	281	20	27	221	240

tr = trace

Table 9b. Geochemistry of the felsic volcanic rocks of the Eden Lake Complex

	THO3-8C	THO3-8D	THO3-8F
SiO_2	76.41	65.41	62.21
TiO_2	0.07	0.64	0.54
Al_2O_3	14.75	16.19	16.17
Fe_2O_3	0.87	4.77	4.98
MgO	0.34	2.56	2.29
MnO	0.02	0.09	0.12
CaO	0.39	0.57	1.35
Na_2O	7.86	5.1	4.97
K_2O	0.06	1.55	1.68
P_2O_5	0.07	0.22	0.19
LOI	0.38	1.85	6.15
TOTAL	101.23	98.95	100.64
Rb	tr	38	52
Sr	68	205	287
Y	21	14	212
Zr	42	220	262
Nb	23	9	8
Ba	19	457	574
Ni	3	2	2
Cr	31	25	22
V	1	66	65

tr = trace

correlation between Fe and Ti and Fe and V (Fig. 68b,g) suggests that titanomagnetite was also an important fractionating phase.

Black Brook Pluton

Field relationships. The Black Brook Pluton is about 8 km^2 in surface exposure and lies about 4 km north of the Chedabucto Fault and 10 km east of Eden Lake in the southern Antigonish Highlands (Fig. 3). Its contact with the surrounding rocks is not exposed. However, it apparently intrudes the late Precambrian Keppoch Formation (Map 1749A) and is intruded by the Devono-Carboniferous Gunn Lake granite to the south.

Age. No contact relationships with the host rocks are exposed. $^{40}Ar/^{39}Ar$ geochronological data on hornblendes (Keppie et al., 1990) yield a 602.5 ± 3.2 Ma plateau age. The strong lithological comparison with the Greendale and Eden Lake complexes suggests an important ca. 620-600 Ma period of appinitic magnetism in the Antigonish Highlands.

Field description. The complex consists of medium to fine grained equigranular hornblende and plagioclase with minor hornblende pegmatites (in pods up to 1 m in diameter) and quartzofeldspathic veinlets and pods.

Petrography. The equigranular rocks consist of prismatic relatively fresh hornblende up to 5 mm in length and plagioclase which is altered to albite, epidote and sericite ± calcite. Opaques and sphene are interstitial minerals. Pegmatites consist of coarse euhedral prismatic hornblende and plagioclase up to 8 cm in length. Quartzofeldspathic rocks are

relatively fine grained and consist almost entirely of quartz and sericitized plagioclase with minor intensely altered microcline.

Geochemistry. The Black Brook Pluton shows a relatively homogeneous chemistry (Table 10). Samples from the Black Brook Pluton are generally higher in Mg and Cr and lower in Al, Fe and in FeO/MgO ratio than either the Greendale or Eden Lake complexes (Fig. 68). This suggests that they are less fractionated. The range in Cr suggests that fractionation of olivine, pyroxene (at depth) or hornblende has occurred (Fig. 68f). The relatively high V (Fig. 68g) reflects the higher modal abundance of magnetite in these rocks.

Discussion

The broad lithological and chemical similarities between the Eden Lake, Black Brook and Greendale complexes together with their approximate contemporaneity suggest an important late Precambrian mafic to intermediate intrusive event occurred in the Antigonish Highlands. The spatial association of these rocks with major bounding faults of the highlands (i.e. the Eden Lake Complex and Black Brook Pluton with the Chedabucto Fault, and the Greendale Complex with the Hollow Fault) and the absence of appinitic gabbros elsewhere in the Antigonish Highlands suggest that the gabbroic magma used these faults as conduits enroute toward the surface. This in turn provides evidence that the Hollow and Chedabucto faults were deep crustal lineaments in the late Precambrian.

Furthermore, this intrusive event may not be confined to the Antigonish Highlands. Similar late Precambrian intrusive rocks have been recognized in the Cobequid Highlands adjacent to the Chedabucto Fault (e.g. the Frog Lake Complex, see Donohoe and Wallace, 1985; Pe-Piper and Piper, 1987).

According to Bowes and MacArthur (1976) the formation of appinites is favoured by high $p(H_2O)$ in magmas of mafic composition. Under these conditions, the stability field of hornblende is greatly increased at the expense of olivine and pyroxene. These conditions are most likely to be achieved at the roofs of magma chambers and/or adjacent to structural traps (e.g. Bowes and MacArthur, 1976).

Although the Eden Lake Complex and Black Brook Pluton strongly resemble the Greendale Complex, there are some notable differences. For example, the coarse hornblende pegmatites, lamprophyres, felsic "sweats", xenoliths and roof pendants are abundant in the Greendale Complex but are not as extensively developed in the Eden Lake and Black Brook complexes. These rock types are generally best developed near the roof of the magma chamber (e.g. Bowes and MacArthur, 1976). Thus the Greendale Complex may be exposed near the roof of the magma chamber and therefore is more typically appinitic, whereas the Eden Lake Complex and Black Brook Pluton are exposed at lower erosional levels and therefore display less pronounced appinitic tendencies.

Table 10. Geochemistry of the Black Brook appinitic complex

	BB1	BB2	BB3	BB4	BB5	596A	593A	59-3B1
SiO_2	51.21	49.59	51.41	51.24	51.53	50.70	45.50	45.50
TiO_2	1.21	1.01	0.98	1.17	1.07	1.05	1.43	0.81
Al_2O_3	15.55	16.92	16.38	15.58	15.42	15.10	16.00	16.60
Fe_2O_3	8.83	8.90	8.34	8.75	8.51	9.54	14.80	9.55
MgO	6.87	7.41	6.59	7.05	6.72	7.45	6.87	8.32
MnO	0.20	0.19	0.20	0.22	0.23	0.23	0.25	0.30
CaO	9.45	8.67	9.02	8.77	9.45	9.13	7.71	13.30
Na_2O	2.83	2.60	3.05	2.94	3.70	3.42	2.90	1.69
K_2O	1.45	1.80	1.67	1.73	0.61	1.10	1.41	0.80
P_2O_5	0.35	0.28	0.26	0.26	0.23	0.23	0.15	0.10
LOI	1.40	1.90	1.80	1.70	1.90	1.70	2.54	2.31
TOTAL	99.35	99.28	99.68	99.4	99.38	99.8	99.7	99.3
Rb	44	58	55	60	12	n.d.	n.d.	n.d.
Sr	446	467	445	425	530	n.d.	n.d.	n.d.
Y	44	19	19	22	19	13	n.d.	7
Zr	94	97	106	86	87	n.d.	n.d.	n.d.
Nb	10	9	9	10	9	4	2	1
Ba	501	544	531	548	216	n.d.	n.d.	n.d.
Ni	43	54	46	44	37	66	56	62
Cr	359	229	295	398	330	260	40	460
V	232	219	206	251	249	n.d.	n.d.	n.d.
Sc	45	35	35	n.d.	n.d.	n.d.	n.d.	n.d.

n.d. = not determined

Alaskite

Field relationships. An alaskite stock 600 m in width occurs 600 m northeast of the Greendale Complex in the northern Antigonish Highlands (Map 1749A; Fig. 3). The pluton is well exposed along the shoreline of the Northumberland Strait. It intrudes the Morar Brook Formation of the Georgeville Group. Smaller bodies of alaskite, associated pegmatites and numerous thin veins occur generally within 1 km of the main body. The intrusive contact is exposed in several localities. The host rock in the vicinity of these contacts is commonly "spotted" (presumably due to the incipient growth of a porphyroblast) and is extensively silicified over a distance of 300 m from the contact. The alaskite is cut by abundant mafic dykes of probable Carboniferous age.

Field description. The alaskite is buff to pale grey, fine grained, equigranular and consists almost completely of quartz, microcline and plagioclase. Mafic minerals are virtually absent except in the vicinity of the contact with the host rock where hornblende, generally elongated perpendicular to the contact, commonly occurs. Amazonite is easily distinguished by its emerald green colour and occurs in discrete pods and veinlets.

The pegmatites consist almost entirely of quartz and microcline. The quartz forms large euhedral crystals up to 70 cm in width and 80 cm in length. Microcline is typically pale orange and forms very large crystals (of undetermined size) that enclose the quartz crystals. Amazonite occurs in isolated pods and veinlets. Accessory pyrite, galena, molybdenite and uranium minerals also occur.

Petrography. The fine grained alaskite consists of quartz, microcline and perthite in graphic intergrowth, minor albite and accessory biotite, muscovite and pyrite. Feldspars commonly display extensive deuteric alteration.

Age. A maximum age is provided by the contact relationships with the host rock. Thin veins of alaskite intrude the country rocks and clearly postdate the late Precambrian deformation in the host rock. The alaskite has been dated by the Rb/Sr whole rock method (R.F. Cormier, pers. comm., 1985), yielding an isochron of 535 ± 13 Ma and an abnormally high initial $^{87}Sr/^{86}Sr$ ratio (0.740). On the other hand, a K/Ar age on muscovite (K = 7.83%) from a pegmatite yielded an age of 604 ± 14 Ma (R.K. Wanless, pers. comm., 1978). It is suggested that the deuteric alteration has significantly modified the Rb/Sr isochron which may be close to the time of this alteration. The significance of the Rb/Sr age is uncertain.

Geochemistry. The alaskite is rich in Si, Na, K and Rb and is very poor in Mg, Fe, Ca, Ti, P and Sr (Murphy, 1982). Murphy (1982) has postulated that the alaskite was derived from the Greendale Complex by the partitioning of Si and the alkalies into H_2O-rich magmas. This would account for the extremely felsic nature of the stock. The presence of amazonite feldspar suggests an oxidized environment during crystallization (Foord and Martin, 1979) although the presence of some sulphides indicates an increase in the fugacity of sulphur during cooling (Stevenson and Martin, 1986).

Precambrian or Cambrian plutons

James River Pluton

Field relationships. A granitic body 3 km in diameter occurs in the James River area (Fig. 3 and 10). The granite intrudes the James River Formation of the Georgeville Group, and apparently truncates structural trends in the Georgeville Group (Fig. 28). Devono-Carboniferous rocks unconformably overlie the pluton to the south (Boehner and Giles, 1982).

Field description. The granite is pink to red, medium to fine grained, equigranular, massive, unfoliated and consists almost entirely of quartz, plagioclase and microcline. Mafic minerals generally account for less than one percent of the rock. The pluton is cut by abundant fine grained mafic dykes of unknown age.

Petrography. The granite consists of anhedral grains of quartz, plagioclase and perthitic feldspar up to 3 mm in size displaying micrographic intergrowth and allotriomorphic textures. Quartz grains show undulose extinction and commonly display optical continuity with adjacent quartz grains. The feldspars are extensively altered: plagioclase alters to albite, epidote, calcite and sericite, and perthite alters to sericite. Primary muscovite is absent and biotite is rare, although minor chlorite pseudomorphs after biotite occur.

Age. A lower age limit is given by the contact relationships which suggest that the pluton postdates the late Precambrian deformation of the Georgeville Group. The upper limit is

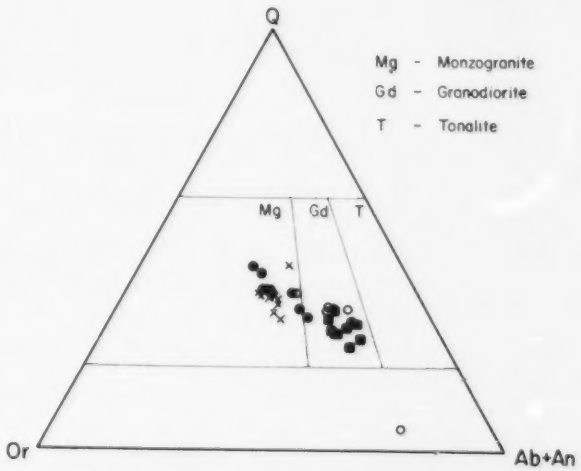


Figure 69. Normative Q-Or-(Ab+An) diagram for the granitic rocks of the Antigonish Highlands. Crosses-James River Pluton, open circles-Ohio Pluton, closed squares-Indian Lake Pluton, closed circles-Gunn Lake. Fields after Streckeisen (1976).

given by the unconformable relationship of the pluton with the overlying Late Devonian-Carboniferous rocks to the south. A preliminary Rb-Sr whole rock isochron by R.F. Cormier (pers. comm., 1984) has yielded a Middle to Late Cambrian age that shows much scatter, probably due to intense alteration. The Middle to Late Cambrian Rb-Sr date, however, is typical of "Avalonian" granites (Cormier, 1979). From the above evidence, it is suggested that the James River Pluton is either latest Precambrian or Cambrian in age. U-Pb analysis on monazite and zircon is currently in progress.

Geochemistry. The James River rocks are high in SiO_2, ranging from 73 to 77 weight percent on a volatile-free basis. They are corundum normative, i.e. peraluminous (Table 11), and plot in the monzogranite field on the Q-Or-(Ab+An) normative diagram (Fig. 69). It is not certain whether the peraluminous chemistry is a primary magmatic characteristic or is due to the loss of alkalies during alteration. The rocks show limited ranges in K_2O, Na_2O and Al_2O_3 (Fig. 70a,b,c), are high in Fe_2O_3t (3.8 to 6.1 weight percent, Fig. 70f) but contain very low concentrations of CaO, MgO, TiO_2 and P_2O_5 (Fig. 70e,g; Table 11).

Trace elements show wide ranges in concentrations. For example, Rb varies from 60 to 160 ppm, and Zr from 80 to 170 ppm (Fig. 71a). The K/Rb ratio varies from 200 to 500 and decreases with increasing Rb (Fig. 71b). Sr is very low in abundance (generally less than 100 ppm) but displays a positive correlation with Ba (Fig. 71c), which varies from 450 to 700 ppm. Zr, Sr, Ba and Rb do not display any apparent correlation with SiO_2 (Fig. 71a,d,e,g). Zn displays a negative

Table 11. Geochemistry of the James River Pluton

	17-60	22-60	48	47-60	44	29-60	44-60	31-60	21-60	30-60
SiO_2	75.12	70.81	78.04	73.70	76.14	75.84	72.84	72.92	73.66	72.92
TiO_2	0.12	0.18	0.16	0.18	0.17	0.15	0.21	0.20	0.19	0.20
Al_2O_3	12.20	12.01	11.06	12.84	13.30	12.74	13.12	12.84	12.64	12.84
FeO^t	4.55	6.05	3.88	3.72	1.58	1.58	3.93	4.54	4.43	4.54
MnO	0.06	0.08	0.03	0.05	0.04	0.03	0.07	0.07	0.07	0.07
MgO	0.71	0.76	0.63	0.70	0.58	0.58	0.68	0.77	0.71	0.77
CaO	0.63	1.14	0.27	0.42	0.24	0.33	0.28	0.42	0.25	0.42
Na_2O	2.65	2.75	3.29	3.08	3.40	3.42	3.40	3.58	3.34	3.58
K_2O	5.31	4.62	3.80	5.01	5.02	4.96	5.07	4.87	4.64	4.87
P_2O_5	0.03	0.04	0.03	0.04	0.03	0.03	0.05	0.02	0.54	0.04
LOI	0.20	1.00	0.20	0.30	0.70	0.60	0.50	0.60	0.20	0.50
TOTAL	100.6	99.4	101.4	100.1	101.2	100.6	100.1	100.9	100.2	100.8
Q	33.8	29.8	39.8	32.3	34.6	34.5	29.5	28.2	31.9	35.1
C	1.0	0.5	1.1	1.7	1.9	1.2	1.7	1.0	1.8	0.9
Or	31.2	27.9	22.3	29.8	29.6	29.5	30.2	28.8	27.6	32.6
Ab	22.3	23.8	27.6	26.2	28.7	29.1	29.0	30.4	28.4	27.0
An	2.9	5.5	1.1	1.8	1.0	1.4	1.1	1.8	1.0	0.4
Hy	8.4	12.0	7.7	7.7	3.8	3.9	8.0	9.2	9.0	3.7
Il	0.2	0.3	0.3	0.3	0.3	0.3	0.4	0.4	0.4	0.2
Ap	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.0
Rb	173	154	116	190	162	161	173	189	152	203
Sr	85	52	59	91	98	48	99	92	61	36
Y	15	18	11	17	14	15	17	20	17	17
Zr	107	144	124	147	137	130	167	164	147	115
Nb	7	7	7	8	8	10	10	9	9	9
Ba	516	545	506	678	666	451	708	661	595	438
Ni	20	16	10	16	17	16	17	21	16	18
Cr	187	236	163	132	45	36	146	157	166	42
V	30	20	20	20	< 10	< 10	20	20	20	< 10
La	35.4	24.7	15.7	27.1	23.3	16.6	27.5	38.2	34.1	24.1
Ce	42.0	27.0	18.0	32.0	24.0	21.0	38.0	50.0	43.0	30.0
Nd	< 5	< 5	< 5	9.0	< 5	< 5	< 5	< 5	< 5	7.0
Sm	2.7	2.2	1.1	1.9	1.6	1.4	2.0	2.8	2.4	1.7
Eu	0.5	0.5	0.3	0.3	0.2	0.2	0.3	0.3	0.3	0.3
Yb	1.1	1.3	1.0	1.6	1.2	1.4	1.7	1.9	1.7	1.8
Lu	0.2	0.3	0.2	0.3	0.3	0.3	0.5	0.5	0.3	0.5
Hf	4	4	4	5	5	5	6	6	3	4
Ta	2	1	1	1	1	2	2	2	2	2
Th	21	13	10	17	9	11	19	13	15	21
Zn	40	50	7	16	34	31	39	52	45	41
Cu	16	9	7	15	1	4	13	12	14	9

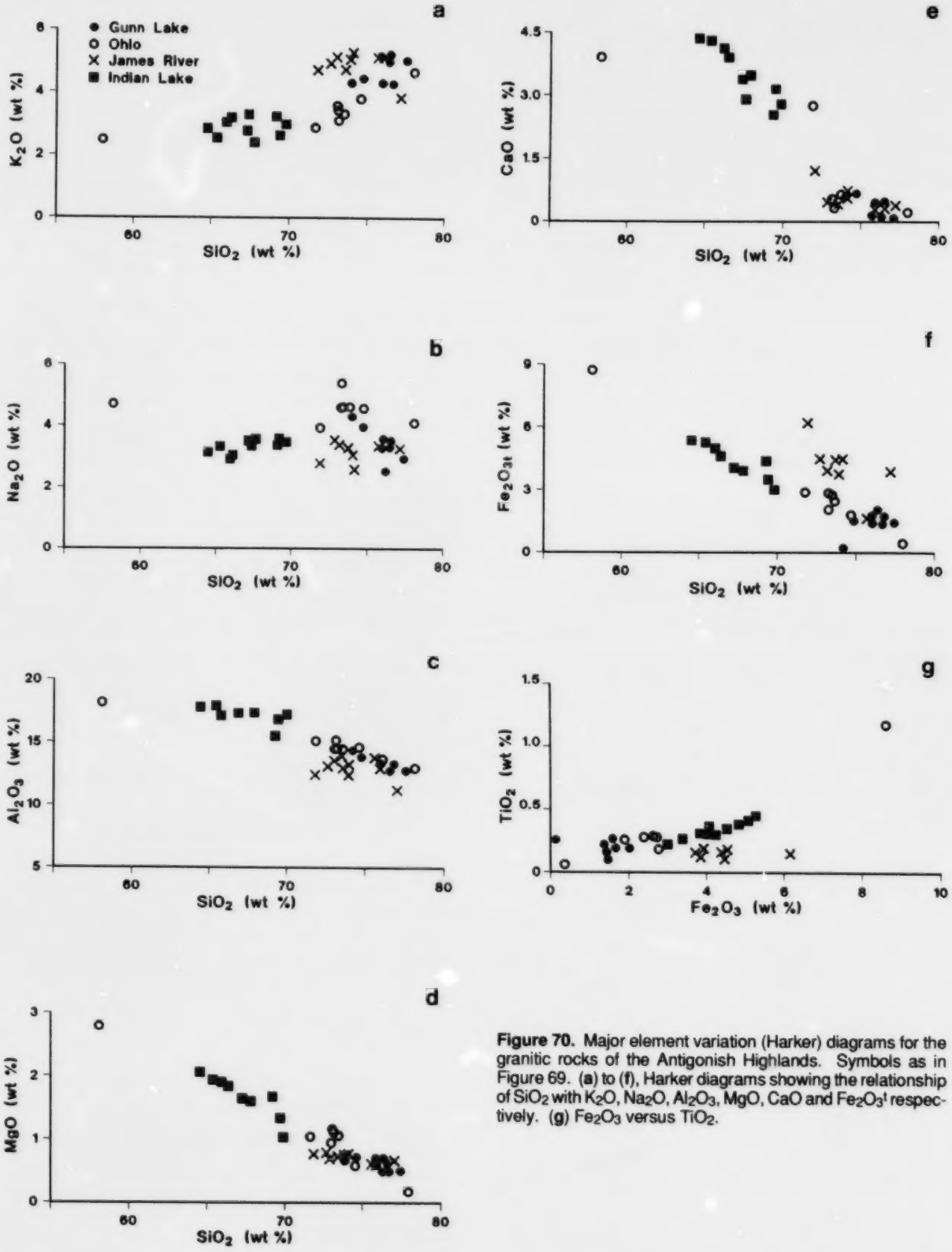
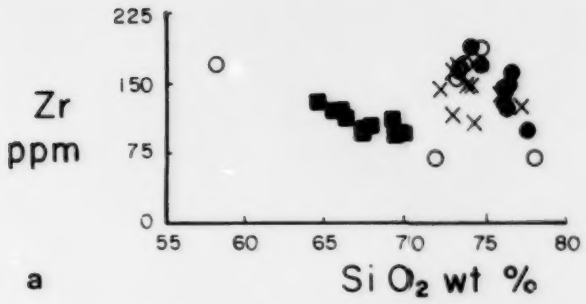
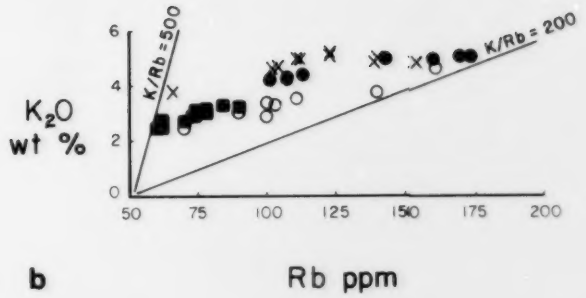
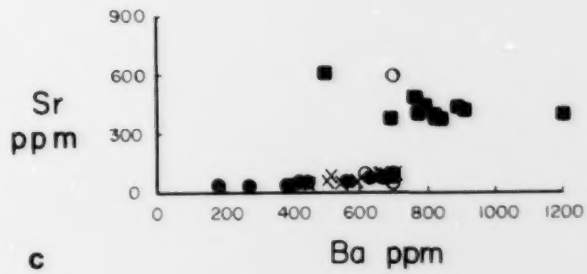
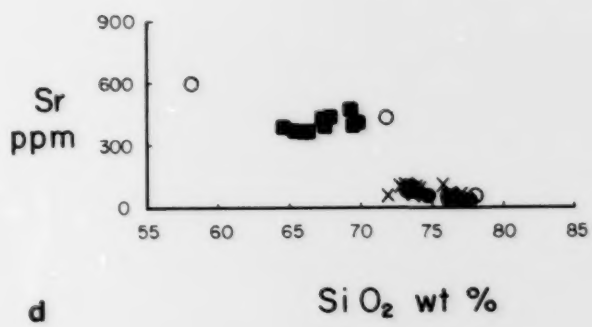


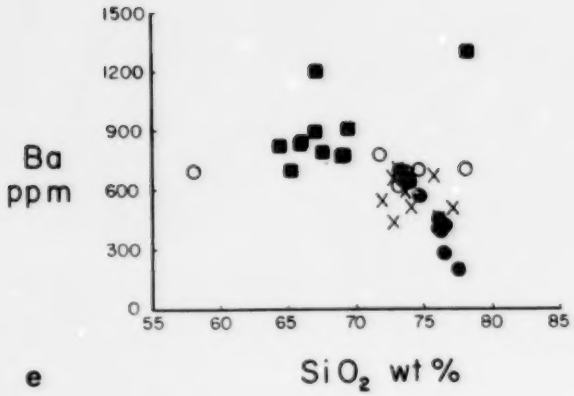
Figure 70. Major element variation (Harker) diagrams for the granitic rocks of the Antigonish Highlands. Symbols as in Figure 69. (a) to (f), Harker diagrams showing the relationship of SiO_2 with K_2O, Na_2O, Al_2O_3, MgO, CaO and Fe_2O_3[t] respectively. (g) Fe_2O_3 versus TiO_2.

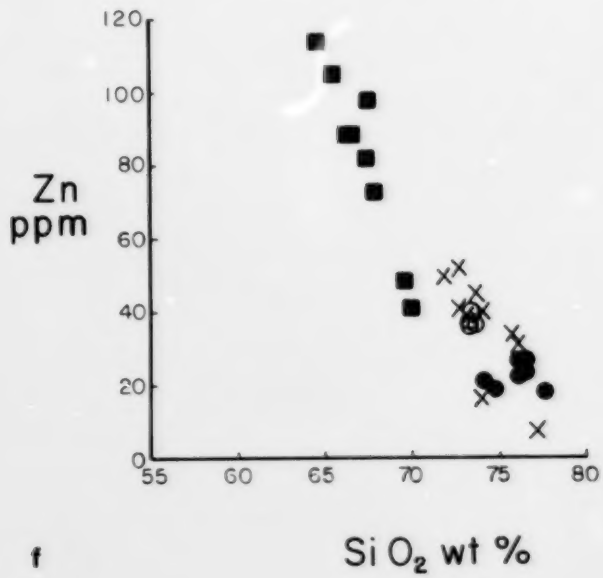












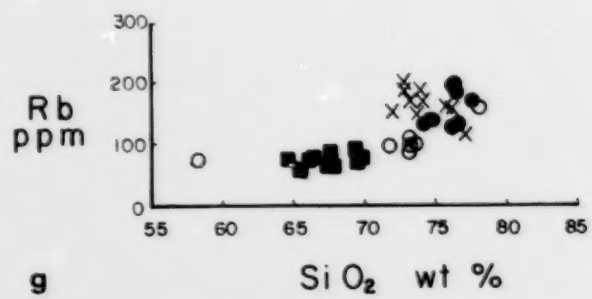


Figure 71. Trace element geochemistry of the granitoid rocks of the Antigonish Highlands. (**a**) Zr versus SiO_2, (**b**) K_2O versus Rb, (**c**) Sr versus Ba, (**d**) Sr versus SiO_2, (**e**) Ba versus SiO_2, (**f**) Zn versus SiO_2, (**g**) Rb versus SiO_2, (**h**) CaO versus Sr, (**i**) Rb versus Ba, (**j**) Th versus Rb and (**k**) Rb versus Sr. Symbols as in Figure 69.

correlation with SiO_2 (Fig. 71f). The granite is LREE enriched with La/Sm ranging from 13 to 15 and La/Yb from 15 to 20 (Fig. 72a).

Tectonic setting. At the time of writing, REE data for the James River Pluton were not available. The low Y and Nb and relatively high Rb is consistent with a syncollisional or postcollisional tectonic setting (Fig. 73). Although it is possible that the abundances of LIL-enriched elements may have been affected by alteration, the multielement profile (Fig. 74a), which shows a depletion in high field strength elements relative to ocean ridge granites (Pearce et al., 1984a) is consistent with a syncollisional or postcollisional setting. Field relationships, which show that the granite is posttectonic with respect to late Precambrian deformation, support the above interpretation.

Ohio Pluton

Field relationships. The Ohio Pluton is well exposed on the Ohio River in the southeastern Antigonish Highlands where it outcrops over an area of 6 km^2 (Fig. 3). Although the contact with the surrounding rocks is not exposed, dykes inferred to be apophyses of the pluton intrude the Keppoch Formation to the west of the pluton and undivided Georgeville sediments to the east. The Ohio Pluton is unconformably overlain by Devono-Carboniferous rocks to the east (see also Boehner and Giles, 1982). Thus the pluton may be considerably larger than what is suggested by the limited surface exposure. Its northern and southern boundaries are faulted contacts.

Age. On the basis of the above field relationships the pluton could be late Precambrian to Devono-Carboniferous in age. Radiometric (Rb-Sr whole rock) data available at present yields an age of 533 ± 19 Ma (Cormier, 1979). This age is typical of "Avalonian" granites in Cape Breton Island (Cormier, 1979) and is similar to that of the James River Pluton. The above data suggests a late Precambrian or Cambrian age of intrusion. U-Pb geochronological analysis is currently in progress.

Field description. The rocks are pale pink, orange and grey medium grained massive equigranular granodiorites. They are composed primarily of quartz, plagioclase and minor potassium feldspar, although some rocks may contain as much as 10 percent mafic minerals (biotite and hornblende). The rocks are relatively homogeneous in composition throughout the pluton. Locally, they have a rusty appearance. Abundant xenoliths of hornblende-bearing diorite and gabbro (similar to the Black Brook appinitic gabbro) occur within the pluton on Ohio River. Aplitic phases are especially common near the margins of the pluton.

Petrography. The rocks are relatively homogeneous in texture and mineralogy, the main variable being the relative proportions of quartz and plagioclase. Quartz and plagioclase display micrographic intergrowth or allotriomorphic textures. Quartz occurs as clusters of discrete optically-continuous grains or as coarse ragged grains that typically display undulose extinction. Plagioclase is albitic in composition and occurs as stubby grains up to 4 mm and as small (i.e. less than 1 mm) ragged crystals. Plagioclase crystals show incipient

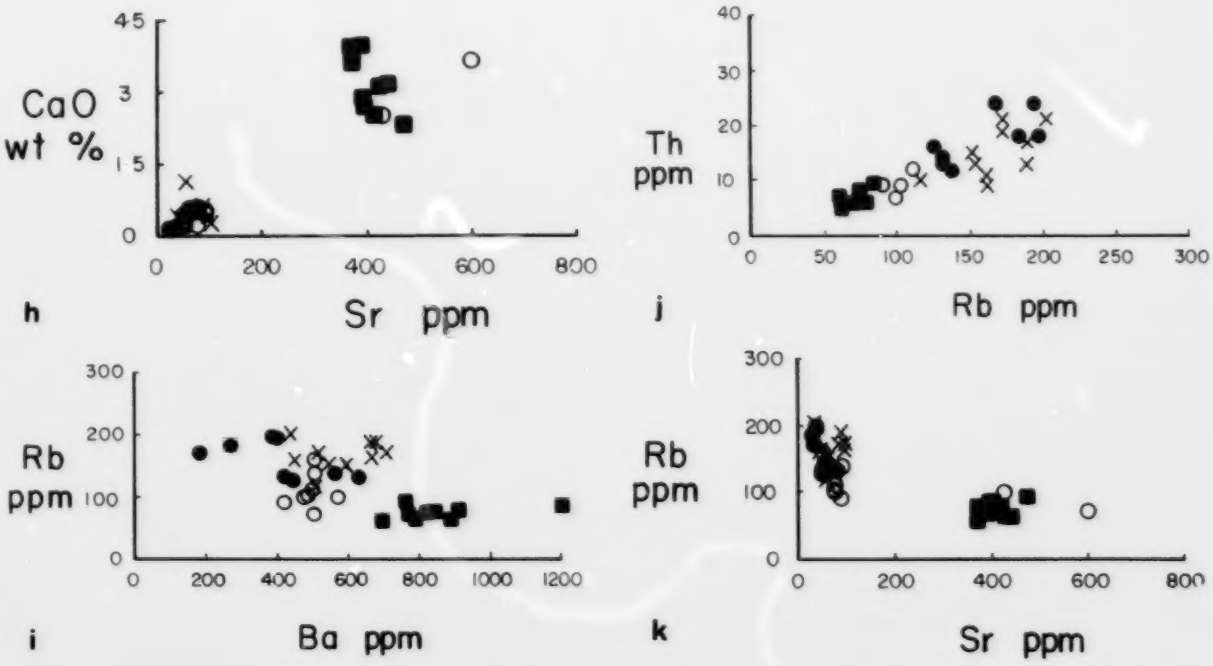


Figure 71. (cont.)

to extensive alteration to sericite and epidote. Potassium feldspar consists of small grains of perthitic orthoclase and displays moderate sericitization. Mafic minerals are rare and constitute less than 10 percent of the rock. They consist of biotite, which is partly to extensively altered to chlorite, and opaques. Opaque minerals consist of magnetite, and in some localities, pyrite. The pyrite is usually altered, giving the rock a locally rusty appearance.

Geochemistry. With the exception of one sample, the Ohio Pluton shows a restricted range in SiO_2 (from 72 to 77 weight percent on a volatile-free basis, Table 12). They are corundum normative (i.e. peraluminous) and plot in the monzogranite and granodiorite fields on the normative Q-Or-(Ab+An) diagram (Fig. 69). The lack of biotite in these rocks implies that the peraluminous character may be due to loss of alkalies during alteration. On Harker diagrams, Na_2O variations are rather erratic (Fig. 70b); however, K_2O displays a positive correlation (Fig. 70a) and Al_2O_3, MgO, Fe_2O_3t, and possibly CaO display negative correlations with SiO_2 (Fig. 70c,d,f,e respectively). Fe_2O_3t displays a positive correlation with TiO_2 (Fig. 70g). For similar SiO_2 concentrations, the Ohio granites contain on average lower K_2O and Fe_2O_3t contents, and higher Al_2O_3, Na_2O and TiO_2 contents in comparison with the James River granites (Fig. 70).

Rb ranges from 60 to 160 ppm and displays a positive correlation with K_2O and SiO_2 (Fig. 71b,g). K/Rb decreases with increasing Rb. Sr shows a moderate decrease with increasing SiO_2 and a positive correlation with CaO (Fig. 71d,h). Ba shows a limited range in concentration and

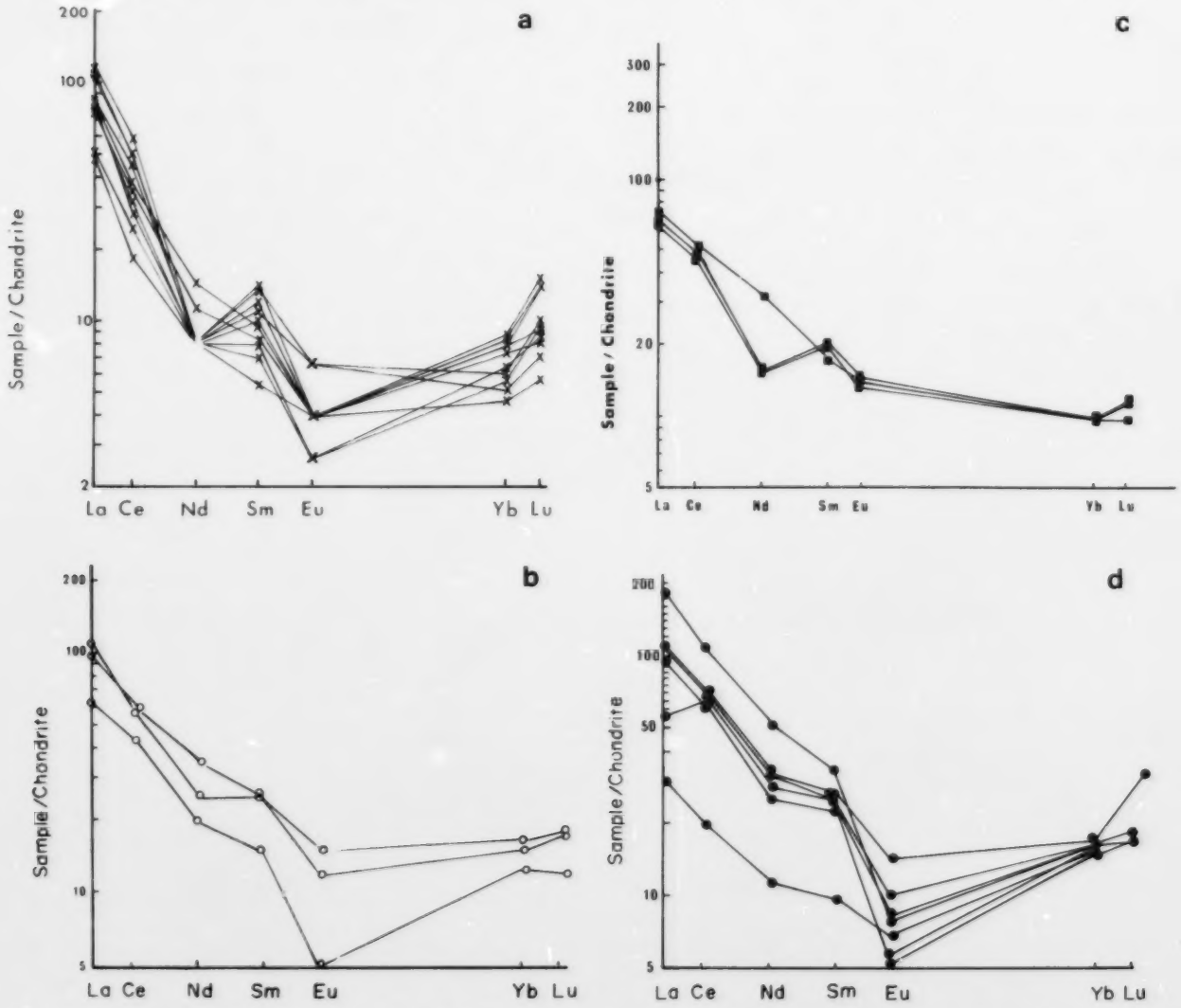


Figure 72. Chondrite-normalized REE diagrams for granitoid rocks in the Antigonish Highlands. (**a**) James River, (**b**) Ohio, (**c**) Indian Lake and (**d**) Gunn Lake plutons. Symbols as in Figure 69.

displays no apparent correlation with Sr, SiO_2 or Rb (Fig. 71c,e,i). Zr varies from 60 to 160 ppm with no apparent correlation with SiO_2 (Fig. 71a). The ranges in concentrations of most trace elements are similar to those in the James River Pluton, with the exception of Ba and Sr which are on average moderately higher in the Ohio Pluton, and Rb, which is lower (Fig. 71e,h,i).

Rare earth elements show strong LREE enrichment with chondrite-normalized La ranging from 65 to 120. Negative europium anomalies are also prominent, although are not sharply defined due to the lack of Gd data (Fig. 72b).

Tectonic setting. The low Y and Nb in all of the Ohio Pluton rocks is consistent with either volcanic arc, syncollisional granites or postcollisional granites and is inconsistent with a within-plate origin (Fig. 73, after Pearce et al., 1984a). Although Rb concentrations are suspect due to alteration effects, their abundances follow expected magmatic trends and their low concentrations in these rocks may be a primary feature. If Rb values reflect primary concentrations, a syncollisional origin is unlikely (Fig. 73). On the multielement variation diagram shown in Fig. 74b (after Pearce et al., 1984a), the granites show less LIL-element enrichment than within-plate or syncollisional granites and contain depleted (i.e. rock/ORG <1) HFS elements. The pattern is generally consistent with a mature volcanic arc or postcollisional setting. The relatively low Sm and Yb suggest a postcollisional setting.

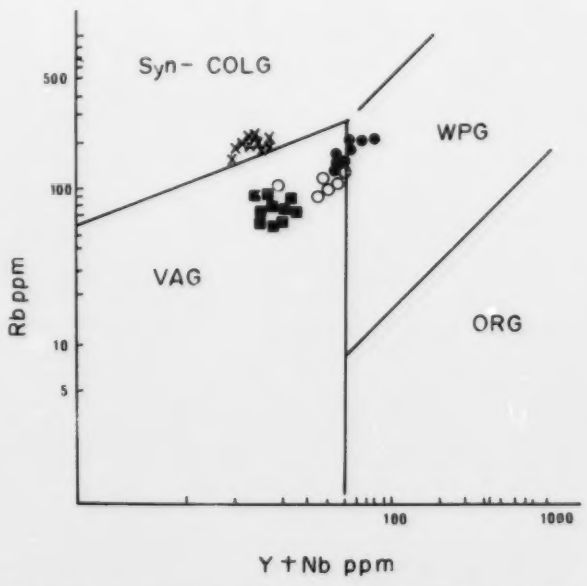


Figure 73. Tectonic setting of the granitoid rocks (after Pearce et al., 1984a). Symbols as in Figure 69. VAG-volcanic ore granite; ORG-ocean ridge granite; WPG-within-plate granite; syn-COLG-syncollisional granite.

Indian Lake Pluton

Field relationships. The Indian Lake Pluton is well exposed on Beaver River and on logging roads in the vicinity of Indian Lake in the central highlands (Map 1749A). Due to the paucity of exposure, especially near the western margin of the pluton, the extent of this pluton is not accurately known. The pluton forms an approximately circular stock with an inferred surface exposure of about 25 km^2. The eastern contact is in part a fault. However, on the northern and western branches of Beaver River, the pluton is inferred to intrude the Keppoch Formation because it contains blocks of rhyolite that were presumably derived from the Keppoch Formation.

Age. The age of this pluton is problematical. On the basis of field relationships, the pluton could be late Precambrian or younger. Wanless et al. (1967) report a 432 ± 18 Ma age (K-Ar on biotite) from a locality included within this pluton. However, the sample was taken adjacent to a major fault and the interpretation of this age is not clear. A Rb-Sr mineral isochron for the Indian Lake Pluton yields an age of 477 ± 21 Ma and an initial ratio of 0.7043 ± 0.0011 (R.F. Cormier, pers. comm., 1986). This age is probably interpreted as a minimum age, probably reset by subsequent tectonothermal events (R.F. Cormier, pers. comm., 1986). 40Ar/39Ar age dating on hornblende and biotite is currently in progress.

Field description. The pluton is a grey massive medium grained equigranular rock. It may be readily distinguished in the field from other granites in the highlands by the relative abundance of mafic minerals (biotite and hornblende) which account for up to 15 modal percent of the rock. On the basis of relative proportions of quartz and feldspar, it is classified as a granodiorite (Streckeisen, 1976). It contains abundant rounded xenoliths of diorite (up to 1 m in diameter).

Petrography. Quartz occurs as anhedral grains up to 4 mm in diameter. Quartz is commonly fractured and displays undulose extinction. Plagioclase occurs as stubby crystals, and has altered cores and relatively fresh rims. The cores are commonly altered to epidote and clay minerals. The rims are albite in composition. Orthoclase forms anhedral grains and displays an incipient alteration to sericite. Biotite is the dominant mafic mineral and is present in all rocks. It forms elongate grains up to 5 mm in length. It is commonly partly to extensively altered to chlorite; however, fresh grains also occur. Hornblende is present in some rocks and occurs as stubby crystals up to 5 mm in length. It is generally partly altered to chlorite.

Geochemistry. In comparison with the James River and Ohio plutons, the Indian Lake Pluton contains higher Al_2O_3, MgO, Fe_2O_3t, CaO and TiO_2, less SiO_2 and similar Na_2O and K_2O (Fig. 70; Table 13). The Indian Lake Pluton ranges in SiO_2 from 64.9 to 70.1 weight percent (on a volatile-free basis). Although the rocks are corundum normative (i.e. peraluminous), the abundance of biotite and hornblende is consistent with a metaluminous chemistry. It is thus possible that these

rocks suffered a loss of alkalies during alteration. On Harker diagrams (Fig. 70), Na_2O shows a moderate positive correlation with SiO_2 and MgO, and Fe_2O_3t and CaO display a strong negative correlation with SiO_2. Al_2O_3 displays a weak negative correlation with SiO_2 and K_2O variations appear to be erratic with respect to SiO_2 (Fig. 70c,a). There is a positive correlation between TiO_2 and Fe_2O_3t (Fig. 70g).

Rb abundances are relatively low (less than 100 ppm). Rb displays a positive correlation with K_2O and decreasing K/Rb with increasing Rb (Fig. 71b). Sr and Ba concentrations are relatively high (Fig. 71c). Sr has a moderate positive correlation with SiO_2 (Fig. 71d) and a negative correlation with CaO (Fig. 71h). Zr and Zn both show a negative correlation with SiO_2 (Fig. 71a,f). In comparison with the James River and Ohio plutons, the Indian Lake Pluton contains lower Rb and Th, and higher Sr, Ba and Zn (Fig. 71).

REE data (Fig. 72c) show moderate LREE enrichment with chondrite-normalized La/Sm of about 4 and La/Yb of about 8. Eu anomalies are not apparent. In comparison to the Ohio Pluton (Fig. 72b), the Indian Lake Pluton contains lower La, Ce, Yb and Lu.

Tectonic setting. The low Y and Nb concentrations are consistent with a volcanic arc or postcollisional tectonic setting (Fig. 73). Rb concentrations, if representative of primary abundances, are too low for a syncollisional origin. LIL-enriched volcanic granites typically show flat, moderately depleted normalized HFS abundances. However, the Indian Lake Pluton is significantly depleted in HFS elements (Fig. 74c) and shows a pronounced negative profile from Ce to Yb. Therefore, a postcollisional setting is favoured.

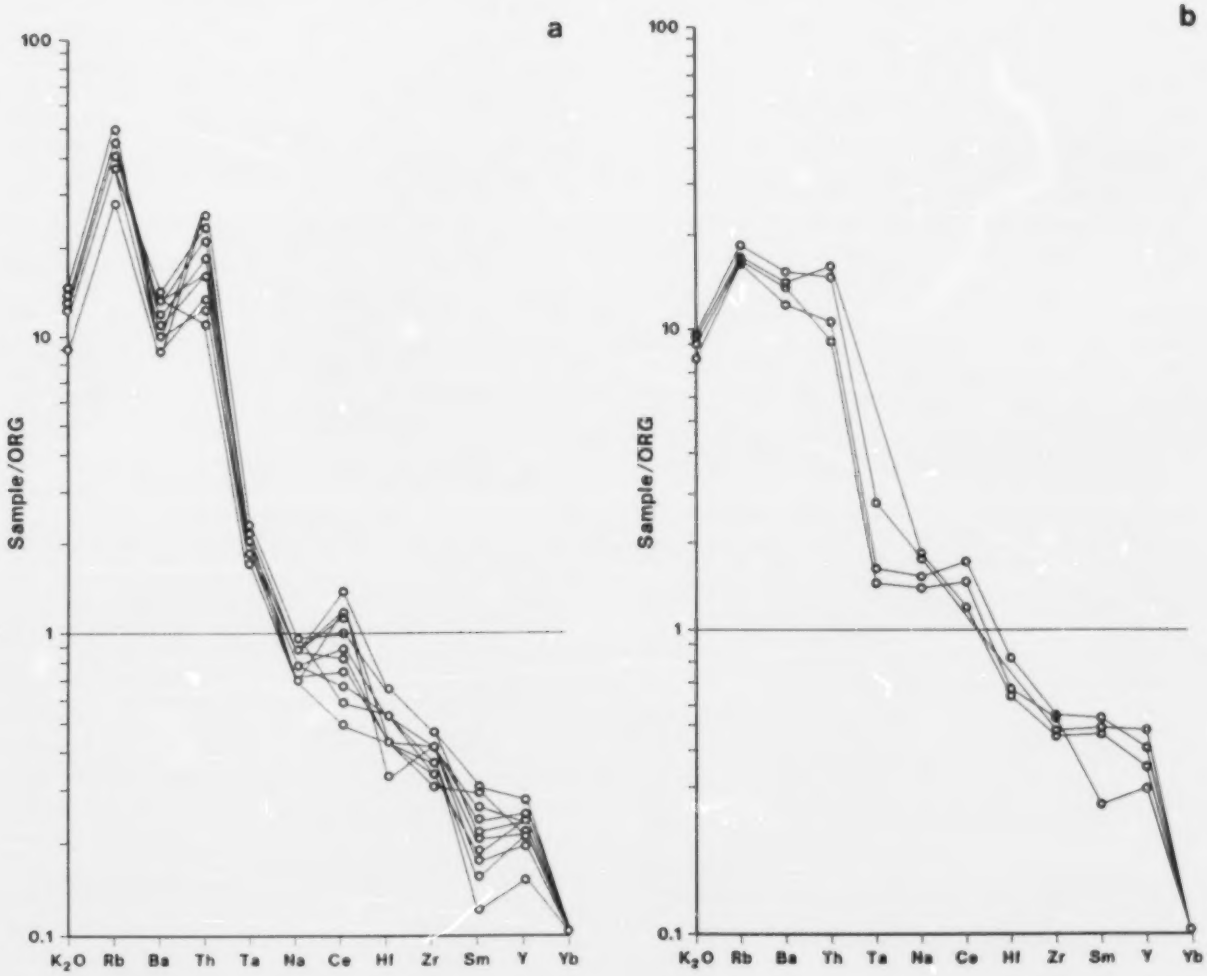


Figure 74. ORG-normalized multielement diagrams (after Pearce et al., 1984a) for (**a**) James River, (**b**) Ohio, (**c**) Indian Lake and (**d**) Gunn Lake plutons.

Discussion

Interpretation of the petrology and tectonic environment is hindered by the lack of reliable geochronological data and by the petrographical evidence of alteration. The available data convincingly rule out a within-plate origin, but volcanic arc, syncollisional or postcollisional tectonic settings are possible. As noted by Colman-Sadd (1982) and Pearce et al. (1984a), it is difficult to characterize the geochemistry of postcollisional granites because: (i) they are generally derived from widely varying sources; and (ii) normal arc magmas may occur in a postcollisional setting. Thus, postcollisional granites may have volcanic arc or syncollisional chemical characteristics.

Geochemical data are broadly consistent with the late Precambrian to Early Cambrian history of the Antigonish Highlands (see section on Precambrian Rocks and Cambro-Ordovician Rocks) and the genesis of the granites appears more likely to be related to the convergent tectonics of the late Precambrian rather than the within-plate extensional tectonics of the Cambrian.

Gabbro

Field relationships. Gabbro bodies occur as stocks 300 to 500 m wide and are spatially associated with the Cambrian basalts of the Arbuckle Brook Formation. The contacts of these stocks are poorly defined. A sill-like body occurs within the Iron Brook Group, near the contact between the Black John and the Little Hollow formations (Map 1749A; Fig. 32 and 37).

Petrography. The gabbros are generally coarse grained and composed mainly of subophitic intergrowths of augite (partly altered to chlorite and actinolite) and saussuritized

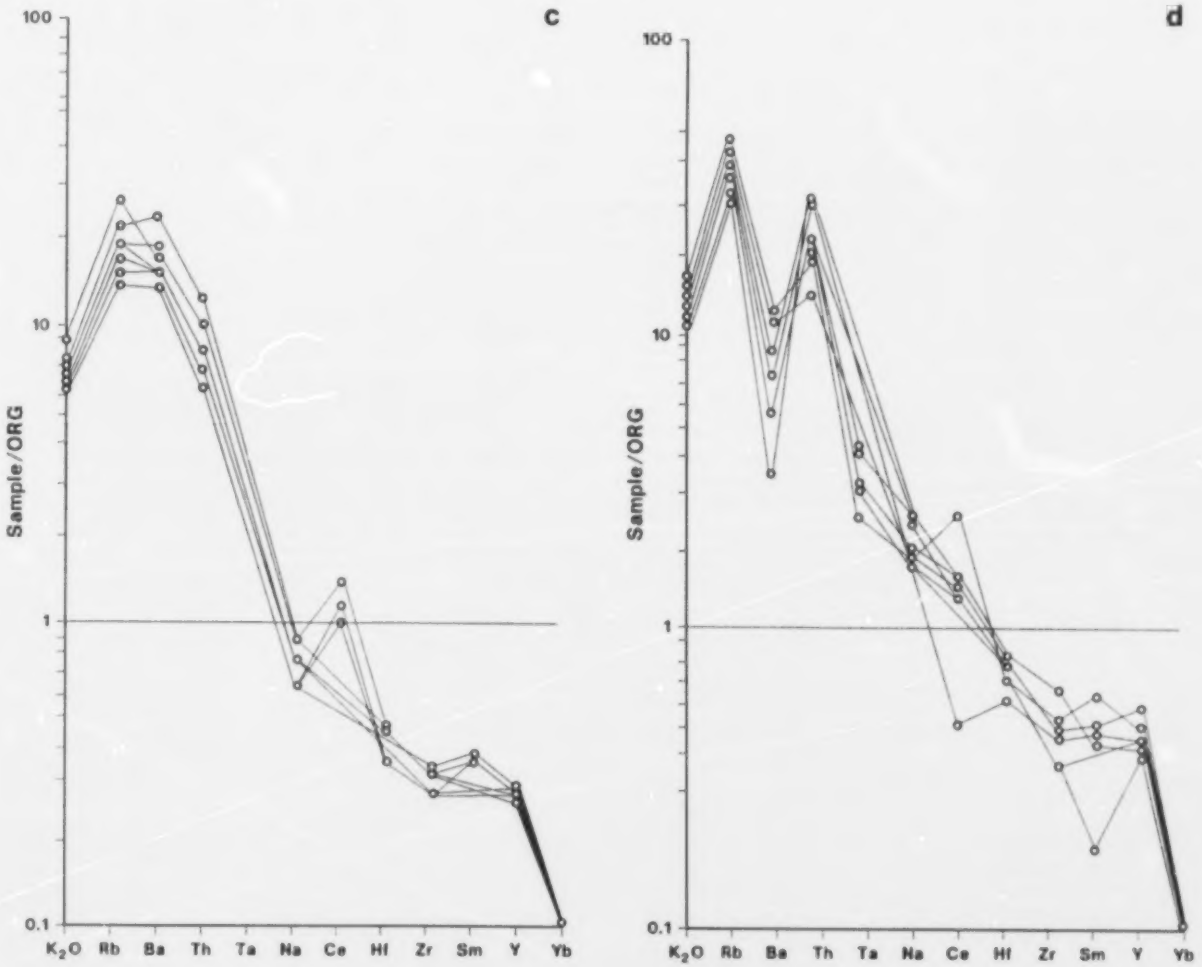


Figure 74. (cont.)

plagioclase, olivine locally pseudomorphed by antigorite with minor chlorite and epidote and accessory magnetite, chromite, magnetite-ilmentite intergrowths and apatite.

Geochemistry. Ten samples of these gabbros were chemically analyzed. Geochemically, the gabbros strongly resemble Cambrian basalts (especially the basalts with low CO_2; compare Tables 4 and 14). This strongly supports the field evidence that these were comagmatic. Discriminant plots (e.g. Fig. 75) indicate that the gabbros are probably alkalic and of a within-plate origin. The range of Ni and Cr in both mafic volcanic rocks and gabbros and the presence of olivine, chromite and augite as anomalous phases in the gabbro suggest fractionation of these phases had an important chemical control. The relatively low Y and P_2O_5 in gabbroic rocks relative to the mafic rocks may to some extent reflect the interstitial intercumulus texture of apatite.

Discussion

Cambrian gabbros are interpreted to be comagmatic with the Cambrian alkalic basalts on the basis of their close spatial and temporal association. Thus, they are probably associated with the late Precambrian-Early Cambrian transpressional stages of the Cadomian Orogeny (Keppie, 1982; Murphy et al., 1985).

Table 12. Geochemistry of the Ohio Pluton

	OHIO-1	OHIO-2	OHIO-3	OHIO-4	59-1C	50-1BD	60-1C	60A-1G
SiO_2	73.00	72.80	73.00	73.30	77.90	71.20	56.50	73.60
TiO_2	0.33	0.30	0.31	0.32	0.06	0.23	1.36	0.23
Al_2O_3	14.00	14.46	13.99	13.87	12.40	14.60	17.00	13.60
FeO[t]	2.68	1.91	2.70	2.39	0.34	2.76	8.44	1.78
MnO	0.06	0.07	0.06	0.06	0.04	0.11	0.14	0.05
MgO	1.09	0.92	1.14	1.03	0.17	1.01	2.72	0.59
CaO	0.28	0.47	0.26	0.58	0.15	2.45	3.51	0.54
Na_2O	4.60	5.39	4.58	4.58	4.09	3.90	4.65	4.50
K_2O	3.37	3.04	3.56	3.29	6.64	2.82	2.45	3.71
P_2O_5	0.10	0.07	0.09	0.09	0.02	0.00	0.44	0.05
LOI	0.60	0.80	1.00	1.20	0.50	0.90	2.40	1.10
TOTAL	100.2	100.3	100.7	100.7	100.4	100.3	99.8	99.9
Q	29.8	26.6	29.0	30.1	35.7	29.6	4.4	30.6
C	2.5	1.6	2.4	1.9	0.4	0.9	1.4	1.3
Or	20.1	18.1	21.2	19.6	27.5	16.9	15.0	22.3
Ab	39.2	46.0	39.0	39.0	34.7	33.4	40.8	38.7
An	0.7	1.9	0.7	2.3	0.6	11.6	15.1	2.4
Hy	6.7	5.1	6.9	6.1	1.0	7.0	19.5	4.2
Il	0.6	0.6	0.6	0.6	0.1	0.4	2.7	0.4
Ap	0.2	0.2	0.2	0.2	0.0	0.2	1.1	0.1
Rb	100	90	111	103	160	100	70	140
Sr	75	90	75	76	50	430	600	90
Y	30	29	23	33	31	17	31	32
Zr	167	156	163	171	70	70	170	190
Nb	12	12	14	13	14	7	14	9
Ba	672	618	697	685	700	770	700	700
Ni	13	10	6	11	n.d.	4	n.d.	n.d.
Cr	25	25	25	29	10	10	30	10
V	11	10	13	10	n.d.	n.d.	n.d.	n.d.
La	33.1	32.1	20.0	n.d.	n.d.	n.d.	n.d.	n.d.
Ce	50.0	52.0	38.0	n.d.	n.d.	n.d.	n.d.	n.d.
Nd	16.0	22.0	13.0	n.d.	n.d.	n.d.	n.d.	n.d.
Sm	4.5	4.7	2.7	n.d.	n.d.	n.d.	n.d.	n.d.
Eu	0.8	1.1	0.2	n.d.	n.d.	n.d.	n.d.	n.d.
Yb	3.1	3.3	2.5	n.d.	n.d.	n.d.	n.d.	n.d.
Lu	0.6	0.6	0.4	n.d.	n.d.	n.d.	n.d.	n.d.
Hf	6	6	6	n.d.	n.d.	n.d.	n.d.	n.d.
Ta	1	1	2	n.d.	n.d.	n.d.	n.d.	n.d.
Th	7	9	12	9	n.d.	n.d.	n.d.	n.d.
Zn	40	37	36	36	n.d.	n.d.	n.d.	n.d.
Cu	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.

n.d. = not determined

Cambrian or younger plutons

Syenite

Field relationships. Syenite occurs only in the northern Antigonish Highlands. The main syenite pluton is located on Iron Brook near its confluence with Doctors Brook, where it has intruded mafic volcanic rocks of the McDonalds Brook Group (Map 1749A; Fig. 32). The northern contact is exposed in Iron Brook and the syenite is fine grained and pale red up to about 8 cm from the contact. The contact here is concordant with the attitude of the mafic volcanic rocks. A single isolated outcrop of a similar syenite occurs in the Precambrian James River Formation south of Sugarloaf Hill (Map 1749A). Another isolated occurrence was found on the shore of Malignant Cove.

Age. The syenites are Early Cambrian or younger. They intrude only Lower Cambrian rocks, and do not occur elsewhere in the Antigonish Highlands. This suggests that they have a genetic relationship with Cambrian rocks. The occurrence of syenite is compatible with the continental rift-related environment envisaged for Cambrian rocks.

Table 13. Geochemistry of the Indian Lake Pluton

	IL-1	IL-1A	IL-2	IL-3	IL-4	IL-6	IL-7	IL-8	IL-9	578G
SiO_2	66.00	66.10	64.40	64.80	67.60	69.30	70.00	67.90	67.10	68.20
TiO_2	0.45	0.44	0.53	0.48	0.39	0.33	0.26	0.38	0.42	0.36
Al_2O_3	16.50	16.50	17.03	16.79	16.68	15.96	16.54	16.59	16.50	14.80
FeO'	4.86	4.59	5.33	5.05	4.07	3.42	3.01	3.97	4.12	4.22
MnO	0.15	0.14	0.16	0.15	0.19	0.12	0.10	0.14	0.15	0.22
MgO	1.88	1.84	2.04	1.89	1.59	1.31	1.04	1.60	1.63	1.60
CaO	3.71	3.57	3.93	3.88	2.68	2.82	2.50	3.13	3.08	2.26
Na_2O	2.97	3.05	3.13	3.25	3.50	3.58	3.57	3.62	3.56	3.40
K_2O	3.06	3.14	2.90	2.42	3.25	2.70	2.98	2.47	2.74	3.13
P_2O_5	0.21	0.20	0.22	0.21	0.21	0.19	0.19	0.22	0.22	0.16
LOI	0.90	1.00	1.20	1.40	0.70	0.70	0.60	0.90	1.10	1.40
TOTAL	100.7	100.6	100.9	100.3	100.8	100.5	100.8	100.9	100.6	100.0
Q	23.6	23.5	20.7	22.9	24.4	28.4	29.3	25.3	24.8	27.1
C	2.1	2.1	2.1	2.3	3.0	2.5	3.4	2.8	2.6	2.1
Or	18.2	18.7	17.3	14.5	19.3	16.1	17.6	14.5	16.3	18.9
Ab	25.3	26.0	26.7	27.9	29.7	30.5	30.2	30.5	30.4	29.4
An	17.2	16.6	18.2	18.2	12.0	12.8	11.2	14.0	14.0	10.4
Hy	12.3	11.8	13.4	12.8	10.4	8.7	7.3	11.6	10.6	11.0
Il	0.9	0.8	1.0	0.9	0.7	0.6	0.5	0.7	0.8	0.7
Ap	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.4
Rb	74	78	73	60	84	70	78	62	62	90
Sr	370	372	388	371	396	391	414	440	425	470
Y	21	20	24	22	21	18	17	18	20	19
Zr	121	111	129	120	101	96	97	102	96	110
Nb	7	7	8	7	9	8	9	7	8	6
Ba	826	841	822	695	1202	769	908	791	890	760
Ni	2	3	3	6	4	4	4	1	6	2
Cr	27	26	22	23	33	39	33	31	36	10
V	70	68	81	81	40	35	19	39	47	n.d.
La	n.d.	22.6	n.d.	20.2	n.d.	n.d.	24.0	n.d.	n.d.	n.d.
Ce	n.d.	43.0	n.d.	38.0	n.d.	n.d.	46.0	n.d.	n.d.	n.d.
Nd	n.d.	10.0	n.d.	10.0	n.d.	n.d.	20.0	n.d.	n.d.	n.d.
Sm	n.d.	3.8	n.d.	3.7	n.d.	n.d.	3.2	n.d.	n.d.	n.d.
Eu	n.d.	0.9	n.d.	1.1	n.d.	n.d.	1.0	n.d.	n.d.	n.d.
Yb	n.d.	2.0	n.d.	2.0	n.d.	n.d.	2.0	n.d.	n.d.	n.d.
Lu	n.d.	0.4	n.d.	0.4	n.d.	n.d.	0.3	n.d.	n.d.	n.d.
Hf	n.d.	4	n.d.	4	n.d.	n.d.	4	n.d.	n.d.	n.d.
Ta	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Th	8	6	6	7	9	6	6	5	6	n.d.
Zn	88	88	113	104	97	47	40	72	81	n.d.
Cu	45	24	25	3	26	21	27	20	51	n.d.

n.d. = not determined

Table 14. Geochemistry of Cambrian gabbroic rocks

	E16-363D	E16-474	E16-363C	E16-477	E16-393	E16-394	E16-363A	E16-478	E16-471	E16-395
SiO_2	46.1	46.2	47.5	48.0	48.2	48.2	48.3	48.5	49.1	52.4
TiO_2	3.3	2.7	3.1	3.1	2.7	3.7	2.6	2.6	2.1	2.3
Al_2O_3	17.5	17.7	17.9	18.2	18.0	17.7	17.7	17.9	16.6	16.8
FeO[t]	13.3	14.1	12.9	11.4	10.9	11.8	13.2	11.9	12.0	11.0
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	6.1	10.3	6.0	4.9	4.6	5.8	4.2	5.6	6.4	4.2
CaO	8.9	4.4	7.5	8.5	11.2	7.8	6.2	7.1	7.8	6.8
Na_2O	4.2	3.7	4.7	4.1	3.6	3.7	4.4	3.6	3.7	3.9
K_2O	0.1	0.5	0.1	1.4	0.5	1.0	0.8	2.0	1.9	2.2
P_2O_5	0.4	0.4	0.3	0.4	0.3	0.3	0.3	0.4	0.3	0.2
Rb	200	21	7	27	4	74	9	31	51	16
Sr	921	316	1041	215	1328	1115	1541	811	682	720
Y	11	9	8	23	8	19	8	25	19	27
Zr	230	151	227	242	240	249	293	235	161	178
Nb	7	12	36	23	13	16	13	17	17	12

n.d. = not determined

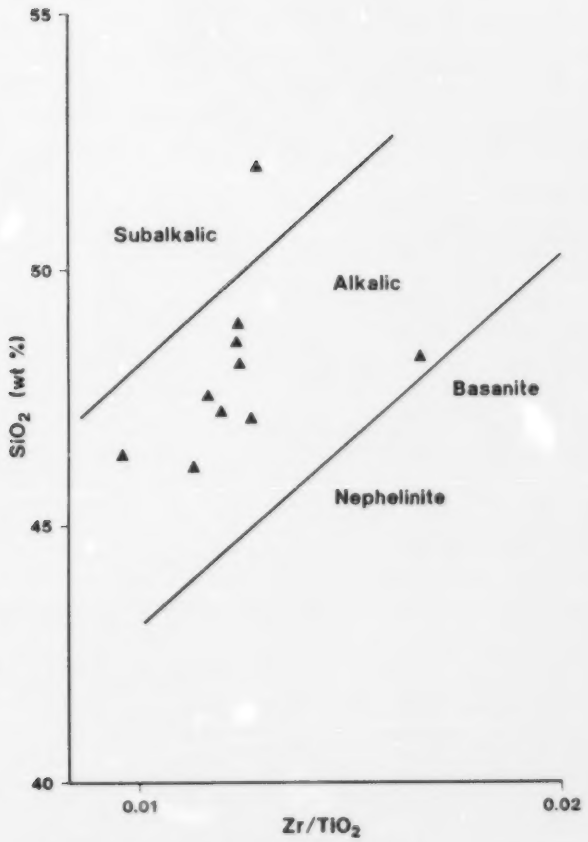


Figure 75. Magmatic affinity of the mafic Cambrian plutons (after Floyd and Winchester, 1975). Closed triangles-gabbro.

Petrography. The syenite has a distinct deep red colour and is intensely altered. Primary mineralogy is almost totally obliterated. The syenite is predominantly composed of feldspar phenocrysts that are strongly sericitized and up to 4 mm in size. The matrix is relatively fine grained and consists of sericite and chlorite.

Geochemistry. Geochemical analyses of the syenite are presented in Table 15. Interpretation of the geochemistry is hindered by petrographical evidence of intense alteration. However, the SiO_2 content confirms their broadly intermediate character. In comparison to the gabbros they contain lower Fe_2O_3[t] and TiO_2, much lower MgO, higher Na_2O and K_2O and much higher Zr, Nb, Y and Nb/Y (Table 15).

If the syenites were derived from the mafic melts, the chemical comparison indicates that this may have involved crystal fractionation of ferromagnesian (pyroxenes, hornblende) and ferrotitaniferous phases (titanomagnetite or ilmenite). However, crystal fractionation processes alone cannot account for the different ratios in incompatible elements between suites. For example, Zr/Nb is generally between 7 and 24 in the gabbros, whereas in the syenites it is between 5 and 7 (excluding sample 502, Table 15). Other processes such as thermogravitational diffusion (e.g. Hildreth, 1981) or liquid-vapour fractionation (e.g. Murphy, 1982) may have occurred.

Granite

Two small granite bodies outcrop near McDonalds Brook within areas underlain by the Iron Brook Group (Fig. 32; Map 1749A). The contacts with the host rock are not exposed but the granite bodies are inferred to intrude Cambrian sediments of the Black John Formation in the MacKinnons Brook area.

Age. On the basis of field relationships the granites may be Cambrian or younger.

Petrography. The granites may be distinguished from all other granites in the Antigonish Highlands by its megacrystic texture. It consists of intensely sericitized potassium feldspar and quartz phenocrysts up to 5 mm in size set in a quartzofeldspathic matrix of quartz, perthite, biotite, chlorite, calcite and magnetite. Sericitization of feldspar is intense in both the megacrysts and the matrix.

Geochemistry. Geochemical analyses (Table 16) are difficult to interpret due to secondary alteration. The SiO_2 content confirms their granitic character. In comparison to the syenites they have higher SiO_2 and lower Al_2O_3 contents; ranges in other elements are broadly similar. A further discussion of the geochemistry is not warranted due to the intense alteration.

Devono-Carboniferous plutons

Gunn Lake Pluton

Field relationships. The Gunn Lake Pluton has a surface area of about 10 km^2 and is situated in the southern highlands about 3 km north of the Chedabucto Fault (Fig. 3). It apparently intrudes the Keppoch Formation to the north and the Ross Brook Formation to the south (Map 1749A).

Age. Only a lower age limit can be established from the above field relationships. Cormier (1979) obtained a 370 ± 15 Ma age (Rb-Sr whole rock)suggesting that the pluton is Late Devonian or Early Carboniferous in age (Palmer, 1983).

Field description. The rocks are red to pink massive, medium to coarse grained equigranular rocks. The pluton consists almost entirely of quartz, plagioclase and microcline in approximately equal amounts. Narrow zones within the pluton display intense shearing and cataclasis, presumably associated with movement on the Chedabucto Fault. The pluton is intruded by mafic dykes of unknown age.

Table 15. Geochemistry of syenites that intrude Cambro-Ordovician rocks in the northern Antigonish Highlands

	E16-341A	E16-341B	E16-341C	E16-407	E16-502
SiO_2	62.05	62.03	62.18	61.22	63.66
TiO_2	0.61	0.41	0.39	0.77	0.76
Al_2O_3	19.64	18.94	17.80	15.87	16.15
FeO^t	6.37	5.94	5.23	5.64	4.89
MgO	1.17	0.43	0.48	2.71	2.17
CaO	0.87	1.94	4.15	5.00	2.92
Na_2O	5.04	6.56	6.49	3.95	4.19
K_2O	4.10	3.62	3.14	4.58	5.04
P_2O_5	0.13	0.12	0.15	0.24	0.22
Rb	143	93	97	232	268
Sr	138	221	294	237	280
Nb	228	230	232	77	31
Y	77	65	69	68	74
Zr	1211	1268	1295	502	514

Table 16. Geochemistry of granites that intrude Cambro-Ordovician rocks

	E16-433	E16-434	E16-435	E16-436	E16-438
SiO_2	68.36	69.12	69.19	70.49	68.86
TiO_2	0.96	0.68	0.64	0.71	0.92
Al_2O_3	14.17	14.21	14.53	14.18	14.27
FeO^t	5.11	4.33	4.13	4.44	5.15
MgO	1.38	0.73	0.75	0.67	1.25
CaO	2.03	1.55	2.00	0.80	1.40
Na_2O	3.71	4.14	4.40	3.66	4.05
K_2O	4.04	5.09	4.19	4.89	3.90
P_2O_5	0.22	0.15	0.16	0.15	0.20

Petrography. On the basis of the relative proportions of quartz, plagioclase and microcline, the rocks are classified as monzogranites (Streckeisen, 1976). Quartz is rounded, up to 4 mm in diameter and displays undulatory extinction. Microcline is perthitic, is commonly altered to sericite and may display micrographic intergrowth with quartz. Plagioclase is generally altered to albite, epidote, calcite and sericite. Primary muscovite is absent. Accessory minerals include calcite, chlorite, sphene and hematite.

Geochemistry. The Gunn Lake Pluton is relatively enriched in SiO_2 (ranging from 74 to 79 weight percent, on a volatile-free basis). The rocks are corundum normative (i.e. peraluminous). On Harker diagrams (Fig. 70) K_2O shows a relatively restricted range in concentration. Na_2O, Al_2O_3 and possibly CaO display negative correlations with SiO_2. CaO concentrations are low (i.e. less than 0.5 weight percent). The ranges in concentrations of the major elements are similar to those of the Ohio and James River plutons.

Rb has a wide range in concentration (between 100 and 200 ppm) and displays a positive correlation with K_2O, with decreasing K/Rb with increasing Rb (Fig. 71b). Sr concentrations are relatively low and display a weak positive correlation with Ba (Fig. 71c). Zr and Ba display negative correlations with SiO_2 (Fig. 71a,e).

Rare earth elements show a wide range in concentration (Fig. 72d) with moderate to strong LREE enrichment and chondrite-normalized La from 40 to 185. All samples show a moderate negative Eu anomaly.

Tectonic setting. The Rb versus Y+Nb discrimination plot is inconclusive. The data plot in both the volcanic arc and within-plate fields. Furthermore, this plot cannot discriminate whether or not these rock types were generated in a postcollisional tectonic setting. The low abundances of the high field strength elements is inconsistent with a within-plate setting (Fig. 74d). If this pluton has a volcanic arc setting then the high LIL concentrations (assuming that this is not an effect of alteration) indicate that the arc was mature. However, mature arcs have higher HFS elements than those shown in Figure 74d. Overall, the above data favour a postcollisional setting and reject a volcanic arc setting.

Discussion

The Late Devonian-Early Carboniferous age of the Gunn Lake Pluton implies that it is broadly contemporaneous with the volcanic rocks of the Middle to Late Devonian McAras Brook Formation, which has been interpreted as a continental rift-related suite associated with the transpressional stages of the Acadian Orogeny (Keppie, 1982; Dostal et al., 1983). The geochemistry of the Gunn Lake Pluton is consistent with this setting and may be related to the docking of the Meguma Terrane to the south of the Chedabucto Fault.

ECONOMIC GEOLOGY

Introduction

Mineral exploration has been active periodically in the Antigonish Highlands for about one hundred years since Fletcher (1887) reported copper, lead and zinc occurrences in the Antigonish Highlands. Detailed descriptions and locations of many occurrences are given in Bourque (1981). Bourque (1981) concluded that significant mineralization in the Antigonish area is, to date, found only in the Carboniferous sediments of the basinal areas within the region: "The Antigonish Highlands . . . show little indication of any substantial mineralization other than Cambro-Ordovician ironstones."

The mineral potential of the Antigonish Highlands is re-evaluated here on the basis of the geological history proposed in this report. The location of the occurrences reported by Wright (1966) and confirmed by Bourque (1981), together with new occurences found during this study, are shown in Figure 10 and Map 1749A.

Mineralization and tectonic setting

The wide variations in tectonic environment exhibited in the Antigonish Highlands implies that many different mineral associations might occur. Figure 76 summarizes the relationship between economic potential and tectonic environment (after Mitchell and Garson, 1981). The volcanic arc/volcanic arc rift tectonic environment for the late Precambrian Keppoch Formation indicates that it is a potential host to a Cu-Au-Mo-Hg or a Kuroko-type deposit such as that exhibited in the late Precambrian Fourchu Group in the Stirling area. The Chisholm Brook Formation with its interarc ocean floor environment may be a suitable area for Cyprus-type or Pt mineral exploration. The Clydesdale seamount may be a favourable environment for Cu and Zn. Paleozoic rocks were deposited in a pull-apart rift environment which often hosts Sn-W-Mo-Ag-Nb deposits.

Mineral occurrences

There appears to be at least five main types of occurrences (Fig. 10):

1) Abundant disseminated pyrite occurs as discontinuous stratabound lenses in interlayered felsic and mafic volcanic rocks of the Keppoch Formation. The best example of this type occurs in the Ireland area on Tozer Road (Murphy, 1984) where the alteration zone is discontinuous (due to incomplete exposure) but is about 300 m thick across strike (Map 1749A). Mapping in the Ireland area shows that pyrite alteration is common on logging roads and brooks (e.g. Callaghan Brook to the south, and as far north as MacIsaacs Brook). Thus, the strike length of this alteration may extend discontinuously for 2 km.

TECTONIC SETTING	MAGMATIC AFFINITY	MINERAL POTENTIAL	N.S. EXAMPLE
Intercontinental rift	Thol	Cu-Fe	T J
Pull-apart rift	Alk-thol	Sn-W-Mo-F-U-Ag-Nb	DC, OS, €
Volcanic arc	Calcalk	Cu-Au-Mo-Hg	P€ Fourchu, Jeffers, Warwick Mtn., Folly River Sch
Volcanic arc rift	Thol	Zn-Cu-Pb-Kuroko type	P€ Fourchu (Stirling)
Interarc ocean floor	Thol	Cu-Pb-Zn Cyprus-type Cr-Ni-Pt	P€ Chisholm Brook Cheticamp
Seamount	Alk-thol	Cu-Zn	P€ Clydesdale

Figure 76. Generalized relationship between tectonic setting and mineral potential.

Table 17. Geochemistry of the Gunn Lake Pluton

	GL-1	GL-2	GL-3	GL-4	GL-5	GL-5a	GL-6	GL-7
SiO_2	76.10	74.20	73.80	75.20	76.10	76.20	77.10	76.30
TiO_2	0.23	0.30	0.30	0.24	0.20	0.20	0.12	0.17
Al_2O_3	13.10	13.30	13.80	12.30	12.80	12.90	12.20	12.80
Fe_2O_3	1.43	1.63	1.75	1.59	1.80	2.02	1.45	1.46
MnO	0.04	0.05	0.05	0.05	0.03	0.04	0.05	0.02
MgO	0.69	0.66	0.64	0.53	0.56	0.66	0.47	0.53
CaO	0.41	0.56	0.51	0.41	0.06	0.11	0.02	0.08
Na_2O	3.68	4.04	4.36	3.60	3.28	2.57	2.93	3.35
K_2O	4.24	4.38	4.30	4.13	5.10	5.07	4.95	4.96
P_2O_5	0.05	0.06	0.06	0.05	0.04	0.04	0.02	0.03
LOI	1.00	1.00	1.00	0.90	0.60	0.90	0.40	0.40
TOTAL	100.9	100.2	100.6	99.5	100.6	100.7	99.8	100.1
Rb	126.0	138.0	132.0	132.0	194.0	198.0	168.0	184.0
Sr	47.0	61.0	79.0	49.0	35.0	32.0	31.0	26.0
Y	24.0	29.0	28.0	27.0	31.0	30.0	32.0	44.0
Zr	143.0	172.0	192.0	163.0	131.0	124.0	98.0	147.0
Nb	18.0	18.0	16.0	17.0	19.0	18.0	25.0	26.0
Ba	449.0	564.0	629.0	420.0	403.0	389.0	183.0	271.0
Ni	7.0	10.0	16.0	15.0	15.0	16.0	15.0	19.0
Cr	26.0	31.0	32.0	54.0	32.0	41.0	34.0	35.0
V	16.0	13.0	14.0	9.0	8.0	10.0	2.0	6.0
La	10.0	38.4	33.1	62.3	n.d.	37.9	n.d.	28.9
Ce	17.0	62.0	54.0	93.0	n.d.	58.0	n.d.	56.0
Nd	7.0	21.0	16.0	31.0	n.d.	20.0	n.d.	18.0
Sm	1.7	4.8	4.2	6.3	n.d.	4.4	n.d.	4.4
Eu	0.4	1.0	0.7	0.6	n.d.	0.3	n.d.	0.5
Yb	3.0	3.4	3.2	3.0	n.d.	3.4	n.d.	5.8
Lu	0.6	0.6	0.6	0.6	n.d.	0.6	n.d.	1.1
Hf	6.0	6.0	7.0	6.0	n.d.	6.0	n.d.	7.0
Ta	2.0	n.d.	n.d.	1.0	n.d.	2.0	n.d.	2.0
Th	16.0	12.0	13.0	14.0	24.0	18.0	24.0	18.0
Zn	23.0	19.0	21.0	24.0	26.0	27.0	18.0	29.0
Cu	<5.0	<5.0	<5.0	<5.0	<5.0	<5.0	<5.0	<5.0

n.d. = not determined

The pyrite is associated with both felsic and mafic rocks and is very fine grained and abundant. The contacts of the zone of alteration with the surrounding host rocks are very abrupt and are parallel to the layering. The felsic rocks are typically bleached to a pale grey colour. Pyrite occurs both as disseminated grains but also within quartz and quartz-carbonate veinlets. More rarely, coarse grained pyrite occurs. The mafic rocks have a waxy appearance and pyrite occurs as very fine disseminated grains. These rocks were analyzed for Au, Ag, Pb, Zn and Cu. Only one sample, to the north of Tozer Road, contains anomalous quantities of Au (150 ppb). Several samples contain anomalous values of Cu.

A detailed analysis of this alteration is beyond the scope of this study. Nevertheless, the occurrence attests to the hitherto unrecognized economic potential of the Keppoch Formation. About 2 km east of this exposure near the end of Tozer Road, Northcote et al. (in press) have documented a vein stockwork in volcanic breccias of the Keppoch Formation that was mined for Pb-Zn-Cu-Au-Ag in the nineteenth century (Fletcher and Faribault, 1887). Although detailed analytical work has yet to be done on these rocks, they indicate a potential for epithermal proximal-distal facies mineralization within the Keppoch Formation.

Other examples of pyritic alteration occur within the Keppoch Formation near the contact between the Moose River and Fraser members (i.e. the contact between subaerial and submarine volcanic rocks) adjacent to highway 347 (Map 1749A).

2) Pyritic zones associated with intrusive "quartz-eye" porphyry occur adjacent to faults, particularly the Browns Mountain Fault and some north- to northwest-trending faults in the Moose River and Ohio areas.

The association of pyritic zones with intrusive rocks along the Browns Mountain Fault may be particularly significant since at least 7 examples have been located over a distance of 45 km. Indeed, it is rare that traverses across the fault do not display some pyritic alteration. One of these localities, the Pleasant Valley area in the northern highlands, was previously reported by Fletcher (1887) and was mined for silver

in the 19th century. Here the pyritization appears to be related to the intrusion of a quartz-eye felsic porphyry close to the fault plane. Another good example occurs in the western Antigonish Highlands (Fig. 10) where sericitized quartz-eye porphyry intrudes Lower Silurian rocks. Other pyritic occurrences along the fault are also closely associated with a similar porphyry. The reported occurrence of Au along this fault in the western highlands could not be confirmed.

Of further interest, perhaps, is that Murphy (1984) suggested that the Browns Mountain Fault may represent the contact between the Precambrian marine (north) and continental (south) domains, and it therefore represents an area of extreme competence contrast and a likely site for faults and magmatism. The age of the dykes is, at present, unknown. The youngest rocks they intrude are the Beechill Cove Formation. If their intrusion is related to felsic magmatism, then the most likely age is Early Silurian or Carboniferous. Chemically they closely resemble Lower Silurian volcanic rocks. The Browns Mountain Fault exerts a significant facies control on Upper Ordovician to Silurian stratigraphy and may have been active at that time. Thus, if the quartz-eye porphyries are Early Silurian, these pyritic zones may be particularly attractive economically in view of the recent discoveries in rocks of a similar age in Newfoundland.

The results of a reconnaissance lithogeochemical survey (Murphy, 1985a) indicate some anomalous concentrations of base and precious metals. Of particular interest is the relatively high base metal concentrations in some of the intrusive quartz-eye porphyries. Values tend to be erratic suggesting that significant mobility may have occurred. Silver values are, however, low.

3) In the Kirkmount area, willemite and native silver are associated with a Zn anomaly in stream sediment surveys reported by Bingley (1977). Occurrences have been documented and described in detail by Sangster (1986) and his observations are summarized below. The occurrence is hosted by veins that cut the James River Formation of the Georgeville Group. Willemite occurs in core drilled by St. Joseph Exploration in 1980, whereas native silver was discovered by Sangster (1986) by detailed surface mapping in a small quarry in the vicinity of the drillhole.

The James River Formation displays moderate hematitic alteration in the vicinity of the drillhole, with fractures and joint surfaces containing hematite, manganiferous oxides and minor barite. Willemite occurs in a complicated network of calcite-bearing brecciated slates and greywackes. Native silver occurs in quartz-carbonate veins.

Sangster (1986) relates the occurrences to the faulting in the vicinity noting that these faults are part of the Cobequid-Chedabucto Fault System which has numerous carbonate-dominant occurrences.

Another example of this type of occurrence may occur in the Georgeville area (Fig. 10). Sphalerite and chalcopyrite have been found in calcite-bearing veins that cut the Morar Brook Formation. The main occurrence is limited to a 1 m wide shear zone. Stream sediment surveys in the vicinity of this occurrence indicate the presence of a Zn anomaly (Bingley, 1977). The veins are spatially associated with local faulting.

These faults are probably genetically related to the Hollow Fault (and therefore the Cobequid-Chedabucto Fault System) which is located adjacent to the present shoreline. Thus the occurrence superficially resembles that at Kirkmount although no willemite has been reported. However, Sangster (1986) noted that the identification of willemite may easily be overlooked.

The occurence of (Pb-rich) amazonite in the nearby Georgeville alaskite and associated pegmatites is significant in that the occurrence of amazonite in pegmatites is often associated with Pb-Zn-Ag ore bodies (e.g. Broken Hill, Plimer, 1976). According to Stevenson and Martin (1986), the formation of amazonite depends on the contribution of Pb from the Pb-Zn-rich host rocks to the magma during crystallization of the feldspar. Therefore, the occurrence of amazonite suggests that the host rocks are anomalously rich in Pb and Zn.

Alternatively, the occurrence may be hydrothermal and related to the nearby intrusion of the Greendale Complex. Chalcopyrite, sphalerite, pyrrhotite and pyrite occur in massive and disseminated forms within the complex adjacent to the contact with the host rock.

4) Two occurrences of copper are found in the College Grant area near Lochaber in the southeastern highlands. Chalcopyrite, malachite, hematite and siderite occur in small veins that cut a dioritic stock and a dark grey shale host of the Early Devonian Knoydart Formation. The age of the stock is uncertain but it is deformed presumably by movement on the adjacent Chedabucto Fault, and therefore probably predates the latest movement on this fault.

To the northeast, pyrite, bornite, sphalerite, pyrite and fluorite were reported in drill core hosted by limestones of the Knoydart Formation (Bourque 1981). According to Bourque (1981) the mineralization is probably epigenetic although its relationship to copper mineralization in the overlying Carboniferous rocks in the Lochaber area is unclear.

5) Minor pyritic occurrences may be found in Silurian rocks west of Eden Lake. Significantly, vertical gradient magnetic maps in this area display a concentric pattern, consistent with the occurrence of a vent complex.

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